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Journal of the Mineralogical Society of America

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STUDIES IN PETROLOGY AND MINERALOGY

Dedicated to

ESPER S. LARSEN, JR.

Professor of Petrography, Harvard University

1923–1949

By Friends and Former Students

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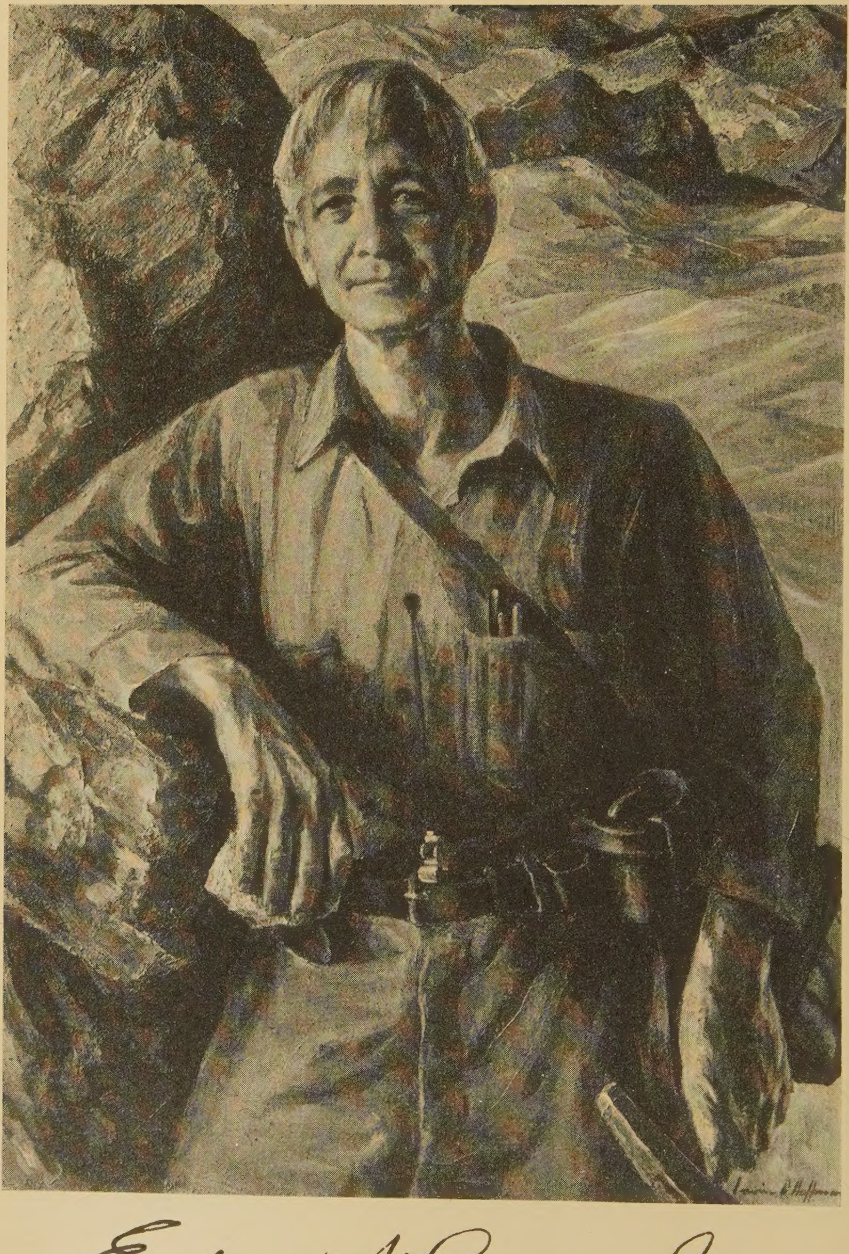

Esper S. Larsen Jr

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1950

ACKNOWLEDGMENT

The reward of a lifetime of research is to be found in the satisfaction one has from having contributed to the sum total of human knowledge; the reward of a lifetime of teaching is to be found in an appreciative and outstanding group of students. Professor Larsen is richly rewarded, for his contributions to the science of geology, petrology and mineralogy are many. The enthusiastic support given in the present volume attests eloquently to the esteem in which he is held by his coworkers in science and his former students.

The publication of this volume has been made possible by contributions from the following:

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The frontispiece is a copy of a portrait of Professor Larsen painted by Irwin D. Hoffman in 1941. The portrait was purchased from funds contributed by thirty-two of Professor Larsen's friends and former students and is now hung in the Department of Mineralogy and Petrography at Harvard University

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THE AMERICAN MINERALOGIST
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Vol. 35 SEPTEMBER-OCTOBER, 1950 Nos. 9 and 10

"THE PROFESSOR" IN HIS LABORATORY AT HARVARD UNIVERSITY.

"THE PROFESSOR"

JOHN C. RABBITT

To those of us who studied under Professor Larsen at Harvard and to many of his colleagues, he is always known as "The Professor." The use of this title in this way sums up the unique esteem in which we all hold him as a man and teacher.

As a teacher, Professor Larsen's methods resulted above all in well-trained students. He insisted on the proper balance of field and laboratory work and he exemplified this in his own scientific career. He had infinite patience with all of his students, the brilliant and the mediocre, the hard worker and the loafer. Steeped as he is in the art and science of petrography it was a privilege to the student to have those individual discussions with him in which his vast knowledge was made so readily available. Nor were these discussions ever on the plane of expert and tyro; rather, they were on a plane of mutual seeking of geologic truth. The students' ideas, ill-conceived as they often were, were considered with respect and thus mutual confidence was attained. In formal lectures, in the field, during demonstrations at the microscope, in informal talks in his office, at the luncheon table, or in his home, this relationship always held. It was a marvel to see him handle a petrographic microscope; he was a natural-born field man and he spared no effort in spending time with his students in the field; informal talks with him were something to look forward to. In all this there was a complete lack of professorial authoritarianism and orthodoxy and yet I never saw anyone who took unfair advantage of this situation. There was no doubt as to who was the leader but it was a leadership which was not based on affectations, histrionics, or a need to uphold a dignity which everyone took for granted.

Coupled with this quality of leadership was Professor Larsen's eminently human regard for his students as individuals. He took an interest in each and everyone of them and, although he might occasionally forget their names, he didn't forget nor neglect their personalities. He was always the logical person to whom personal problems were taken with the assurance that sound advice and practical help would invariably be forthcoming. I have never regarded him as a sentimental man (he is rather a sensitive man); the sound advice and practical help were such as were most useful at the moment. The practical help, in which he was joined by Mrs. Larsen, was the kind that is supplied by "good neighbors" (rare in this modern world) and often resulted in sacrifices and discomforts to the Larsens.

Professor Larsen never enters much into scientific controversy. He has a regard for the other man's point of view and he isn't likely to ascribe an invidious motive to those who disagree with him. Confronted with a correction of some of his early work, his reaction is usually one of pleasure that we are all getting closer to the truth. He can and does, however, defend vigorously theories and work that seem to him to be reasonable but he does this with the least amount of sound and fury and in a manner that is least likely to injure professional feelings. As a result, there are few if any people who have strong feelings of derogation about him or his work. To put it plainly, he is one of the few people I know to whom are directed no unkind words, professional or private. I think this is because of his essential kindliness and because he maintains a young mind. He doesn't cling dogmatically to untenable notions in the scientific field and he keeps himself young in spirit and outlook through association by preference with the younger men, to their and his advantage.

I hope I haven't made Professor Larsen sound like a repository of all virtues. He is, of course, not that. He is, however, a man of tremendous scientific ability and attainment, he was and is a fine teacher, and he is a good man. He manages to be all these things by being just himself.

Professor Larsen retired from teaching at Harvard over a year ago (June 1949) but in his present position as consultant to the Geochemistry and Petrology Branch of the U.S. Geological Survey he is in a sense continuing his career as a teacher in a broader field. He has many scientific "irons in the fire" and they are being brought to the right temper with his customary vigor and vision. He is becoming acquainted with a new group of young professional colleagues and is renewing contacts with old friends in and out of the Survey. We all wish "The Professor" many more years of fruitful work in his chosen field and relaxation with his many friends.

INTRUSION RELATIONS OF BAHIAITE FROM SOUTHERN NORWAY

Tom. F. W. Barth,
Geologisk Museum, Oslo, Norway.

Abstract

Simple holocrystalline dikes of bahiaite (hypersthene and hornblende) cut a small body of massive hornblende norite surrounded by pre-Cambrian agmatitic gneisses. It is believed that super-critical water vapor streaming through cracks in the norite could transform the adjacent rock to bahiaite having a dike-like relation.

Bahiaite is a holocrystalline, massive rock characterized by the combination of hypersthene and hornblende. It was first described by Merrill (1895, p. 658) from Grass Creeks, Montana. Later the same combination was described by Washington (1914, p. 84) from near Maracas, Bahia, Brazil, who writes: "As the type is now shown to be widespread, I venture to suggest the name bahiaite. . . ." The Bahian rock occurs as a mass about 1 km long, 200 meters wide with distinct foliation parallel to the contact of the surrounding pre-Cambrian gneisses.

Bahiaite has been encountered within the large pre-Cambrian gneiss district of Southern Norway (Barth, 1947, p. 47). The rocks of this district are predominantly granitic in composition, but schlieren, patches, and irregular bodies of amphibolite—i.e. metamorphic, schistose combinations of plagioclase and hornblende of uncertain origin—are frequently encountered; in places they are strongly concentrated and dominate the rock composition.

Some of the larger amphibolite bodies represent metamorphic norites and contain locally, small enclaves of non-schistose rocks i.e. massive gabbroidal rocks, which by gradation pass into the surrounding amphibolite of schistose habit. Bahiaites associated with hornblendites occur at several places as special facies of the massive gabbroidal rocks. Thus the field relations indicate that such bahiaites, although completely massive, and without any obvious sign of secondary alteration, are of metamorphic-metasomatic origin. Bahiaite of other relations likewise occurs.

Figure 1 shows a map of a small part of the South-Norwegian gneiss district that was surveyed by me some years ago (Barth, 1945). The *mixed gneisses* represent a heterogeneous migmatic-anatectic rock complex in which amphibolites are present as bands, zones, schlieren, and irregular patches in a granitic matrix. The *hornblende granite* is probably younger than the gneisses, and exhibits a more igneous look, but certain facies of it are just as "gneissic" as the mixed gneisses. The contact be-

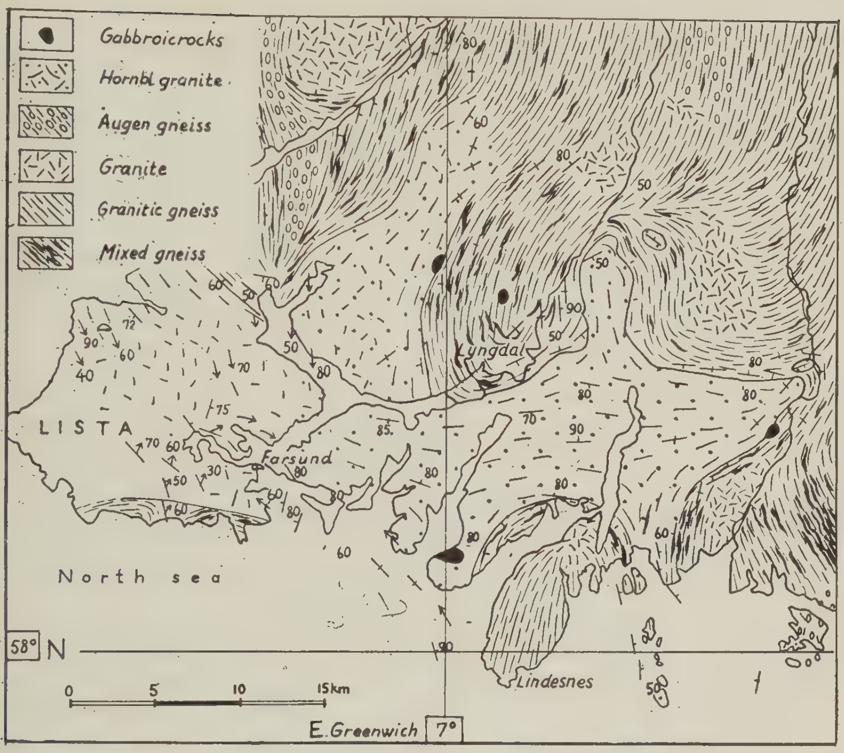

FIG. 1. Geological map of the southern tip of Norway.

tween gneiss and hornblende granite is interesting: As indicated by the map at the end of the long offshoot that extends northward, the contact is transitional and indefinite. Everywhere else the contact is sharp and conformable; both the hornblende granite and the surrounding gneisses are strongly foliated parallel to the common border.

Gabbroic rocks are represented by four small bodies within the map area. They are massive without foliated borders. Bahiaite is associated with the most northern of these bodies and will form the subject of the following description.

The field relations are shown by Fig. 2. A massive body of hornblende norite that is situated right on the border between hornblende granite and mixed gneiss is dissected by a number of small dikes of massive bahiaite. Small patches of hornblende norite, detached from the main body, are scattered around. In this place numerous small, unsystematically distributed, angular inclusions of amphibolite make the mixed gneisses look like agmatite, i.e. a gneissified breccia.

The hornblende norite: Plagioclase forms clear, homogeneous, polygonal crystals, several millimeters long, of composition 57An. Hypersthene forms clusters of equidimensional small crystals; (several tenths of a millimeter long). They are distinctly pleochroic; indices of refraction are $\alpha=1.715$ (red), $\beta=1.727$ (yellow), $\gamma=1.731$ (green,) $(-)2V=54°$; corresponding to 51 Fs (according to the diagram of Poldervaart 1947, p. 167). Hornblende is hypidiomorphic homogeneous, and shows rather large, prismatic crystals, several millimeters long; pleochroism is brown-greenish, brown-light yellow.

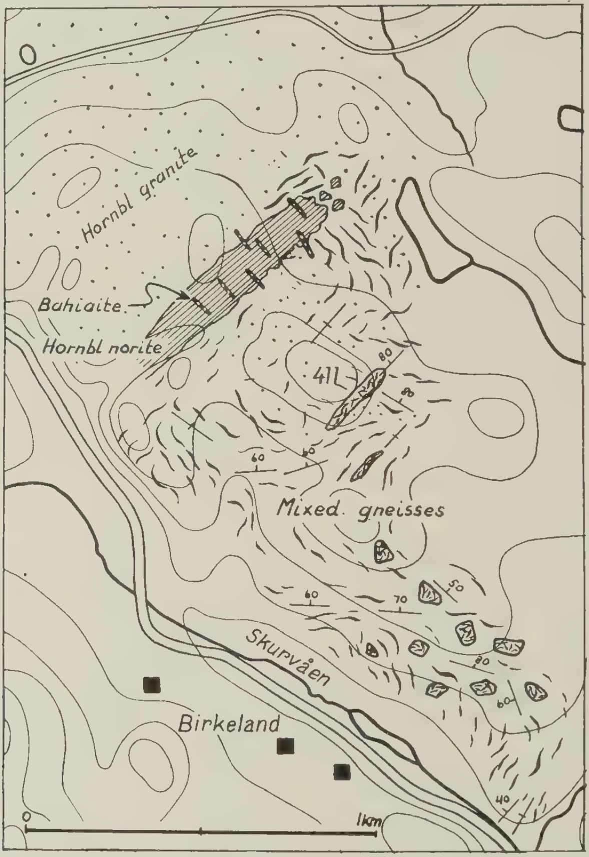

FIG. 2. Geological map of the hornblende-norite—bahiaite occurrence near Lyngdal, Southern Norway.

The bahiaite contains no feldspar. Hypersthene and hornblende are similar to those of the norite. Hypersthene shows $\alpha=1.706$, $\beta=1.718$, $\gamma=1.722$, $(-)2V=58°$ corresponding to 44 Fs.

Chemical analyses of both hornblende norite and bahiaite are entered in Table 1. The simple mineral composition of these two rocks may easily be calculated from the analyses. The quantitative mineral assemblages thus calculated are listed in Table 2.

The compositions of the two hornblendes, that of the norite and that of the bahiaite, as calculated from the corresponding rock analyses are respectively:

$$[Na_{0.8}Ca_{1.9}Mn_{0.1}]_{2.8}[Mg_{2.0}Fe_{1.6}''Ti_{0.6}Fe_{0.5}'''Al_{0.3}]_5[Al_{2.9}Si_{5.1}]_8O_{23}\cdot H_2O \quad \text{and}$$

$$[Na_{0.7}Ca_{1.8}Mn_{0.1}]_{2.6}[Mg_{2.5}Fe_{1.4}''Ti_{0.4}Fe_{0.3}'''Al_{0.4}]_5[Al_{2.3}Si_{5.7}]_8O_{23}\cdot H_2O.$$

The two hornblendes are rather similar in composition. The iron: magnesia relations i.e. $\frac{Fe}{Fe+Mg}$, expressed in terms of molecular percentage of iron are for hypersthene and hornblende respectively: in norite 50 and 49, in bahiaite 44 and 40.

It is worth noticing that in both rocks hornblende is enriched in magnesia relative to the hypersthene with which it is in contact. In hornblende-bearing basalts the opposite relation exists.

Table 1. Chemical Analyses

	1	2	3	2′	3′
SiO_2	44.38	43.57	42.78	41.3	40.6
TiO_2	1.98	2.49	2.95	1.8	2.1
Al_2O_3	15.46	16.33	9.14	18.3	10.2
Fe_2O_3	3.27	3.01	2.79	2.2	2.0
FeO	10.17	14.17	18.94	11.3	14.9
MnO	0.26	0.22	0.33	0.2	0.3
MgO	7.27	7.75	13.64	11.0	20.6
CaO	9.24	8.52	6.24	8.7	6.3
Na_2O	3.31	2.38	1.33	4.4	2.4
K_2O	0.90	0.29	0.21	0.3	0.2
H_2O^+	3.48	0.98	1.04	3.1	3.3
H_2O^-		0.02	0.05		
P_2O_5	0.28	0.22	0.46	0.2	0.3
CO_2		0.27	0.05	0.3	0.1
S		0.22	0.07		
F		0.04			
Cl			0.06		
Total	100.00	100.48	100.08	100.0	100.0

1. Average of six crinanites, after Daly (1933, p. 22).
2. Hornblende norite. Near Lyngdal, Southern Norway (Bruun analyst).
3. Bahiaite. Near Lyngdal, Southern Norway (Bruun analyst).

2′ and 3′. The corresponding rocks in terms of equivalent molecular percentage, or per cent of cations.

The field relations and the similarity of the mineralogy strongly indicate that the two rocks are congenetic. Let us accept without argument that the rocks are magmatic and let us make a routine petrochemical interpretation of the analyses: The composition of the hornblende norite, although unusually high in iron, is not far from that of an olivine basalt; Table 1 demonstrates that it is close to the crinanites or analcime-olivine diabases from Scotland. The composition of the bahiaite is slightly more magnesian and approaches that of an ultramafic rock which may be regarded as an early differentiate, or an accumulative phase of the olivine-diabase magma.

TABLE 2. MOLECULAR NORM AND MODE

Hornblende norite				*Bahiaite*			
			Norms				
		Wo	2.6			Wo	4.0
Or	1.5	Hy	10.6	Or	1.0	Hy	65.8
Ab	22.0	Ol	21.0	Ab	12.0		
An	34.0	Mt	3.3	An	19.0	Mt	3.0
		Il	3.6			Il	4.2
	57.5	Ap	0.6		32.0	Ap	0.8
		Cc	0.6			Cc	0.2
			42.3				78.0
			Modes				
Plagioclase (57An)			35.0				
Hypersthene (51Fs)			26.5	Hypersthene (44Fs)			43.2
Hornblende			34.7	Hornblende			53.6
Ore*			2.8	Ore**			2.2
Apatite and calcite			1.0	Apatite and calcite			1.0
			100.0				100.0

* Composition: $FeTiO_3=1.0$, $Fe_3O_4=1.5$, $FeS=0.3$.

** Composition: $FeTiO_3=1.2$, $Fe_3O_4=0.9$, $FeS=0.1$.

These facts indicate the following intrusive history:

(1) A forceful intrusion of a homogeneous magma highly charged with vapor into a narrow vent (shattering of the surrounding rocks—formation of agmatite).

(2) The congealing at low temperature of this magma to hornblende norite. Normally the magma should have given the combination plagioclase-pyroxene-olivine which corresponds to a much higher temperature.

Water vapor and other gases were responsible for the low temperature. (3) The formation of cross joints and the subsequent intrusion of a hot, plastic stiff mud composed of antecedent crystals of hypersthene and hornblende which had accumulated at the bottom of the reservoir. This last assumption is necessary for two reasons. Bahiaite is an *early* differentiate of the norite and cannot have been intruded as a magma posterior to the formation of norite. A bahiaitic magma, i.e. an ultramafic magma does not exist at a reasonable temperature.

The interpretation does not tally too well with the field relations although it seems to be the only possible one, if we assume an igneous rock formed by crystal fractionation in a magma and subsequent intrusion.

One fact that makes any magmatic interpretation difficult is the isolated location of the small gabbro bodies. Where is the magma reservoir in which the differentiation is supposed to have occurred? There is certainly no geological evidence of it in the neighborhood.

Secondly the intrusion mechanism of the bahiaite: stiff masses of crystals lubricated by thin films of liquid being forced through narrow openings, should have resulted in crushing and grinding of the crystals, friction striae on the walls, and an oriented arrangement of splinters and elongated crystals of hornblende. No such effects can be seen.

It may be necessary, then, to consider a metasomatic mode of origin. The norite body lies on the contact hornblende granite—mixed gneiss. The gneiss in the immediate neighborhood is unusual in that it is developed as an agmatite. The evidence is strong that the place once was a chimney or blow-hole through which vapors and emanations were expelled.

The metasomatic action of these emanations on the existing rocks (for example on a large amphibolitic inclusion) resulted in the formation of the hornblende norite (the temperature was probably slightly higher than that of the metamorphism of the surrounding amphibolites). By further pneumatolytic action the bahiaite dikes were formed according to a mechanism suggested by Bowen and Tuttle in 1949 (p. 459): water vapor charged with various rock-making oxides streaming through a crack in the hot hornblende norite could convert the adjacent rock to bahiaite which might appear to be a dike in the hornblende norite.

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METAMORPHOSED MAFIC DIKES OF THE WOODSVILLE QUADRANGLE, VERMONT AND NEW HAMPSHIRE

Marland P. Billings, *Harvard University, Cambridge, Mass.*
and Walter S. White, *Calumet, Michigan.*

Abstract

Metamorphosed mafic dikes in the Woodsville quadrangle, Vermont and New Hampshire, intrude metasedimentary and metavolcanic rocks ranging from Middle Ordovician to Early Devonian. The mafic dikes are probably Late Devonian. The grade of metamorphism in the metasedimentary and metavolcanic rocks ranges from low-grade (chlorite zone) to high-grade (sillimanite zone). The metamorphosed mafic dikes, many of which were presumably diabases before metamorphism, now show seven principal mineral assemblages as well as several transitional types.

In the chlorite zone the metamorphosed mafic dikes are now characterized by five significant mineralogical assemblages. One of these assemblages, the metadiabases, represent rocks that preserve relics of the original igneous minerals; these rocks did not attain equilibrium in the chlorite zone. A second group, the albite-actinolite-epidote-chlorite rocks, represent a stable assemblage to which no carbon dioxide was added during metamorphism. A third assemblage, the albite-chlorite-epidote-calcite-actinolite schists, represent rocks to which some carbon dioxide was added during metamorphism. A fourth group, the albite-chlorite-calcite schists are rocks to which so much carbon dioxide was introduced that all the lime was utilized to make calcite except for the small amount of lime that went into albite. A fifth assemblage, the albite-chlorite-ankerite schists are rocks into which so much carbon dioxide was introduced that some iron and magnesia as well as all the lime, except that present in the albite, went into carbonate.

Albite-epidote amphibolites formed in a belt lying in the higher grade part of the chlorite zone and in the biotite zone; but there are also many green schists in this same belt. Amphibolites developed in the higher grade part of the garnet zone and in the staurolite and sillimanite zones.

Introduction

Metamorphosed mafic dikes and sills are common throughout eastern Vermont and western New Hampshire. These rocks, as well as the metasedimentary and metavolcanic rocks into which they are intruded, show a great range in the grade of metamorphism. Consequently, an excellent opportunity is available to investigate the mineral assemblages in equilibrium in metamorphosed mafic rocks under a wide range of physical conditions.

The Woodsville quadrangle covers an area of about 210 square miles in east-central Vermont and west-central New Hampshire between north latitudes 44°00′ and 44°15′ and west longitudes 72°00′ and 72°15′. The structure, stratigraphy, and metamorphism of this area has been described elsewhere (White and Billings, 1950). It will suffice here to say that the metasedimentary and metavolcanic rocks range from Middle Ordovician to Early Devonian and are approximately 35,000 to 40,000

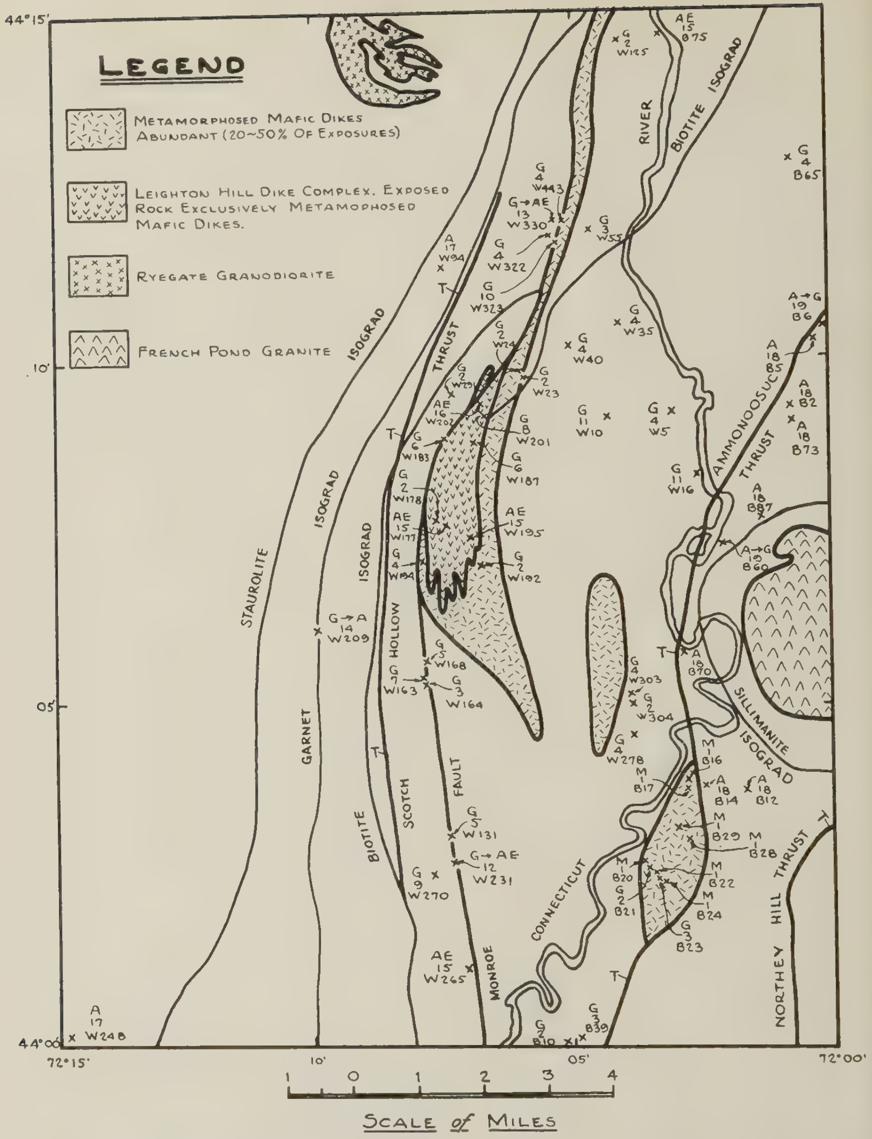

FIG. 1.—*Location of Metamorphosed Mafic Dikes Studied Microscopically in the Woodsville Quadrangle.* Principal faults and isograds are labelled. In the eastern part of quadrangle the biotite, garnet, and staurolite isograds coincide with the Ammonoosuc thrust because the chlorite and staurolite zones have been brought together. Location of each metamorphosed mafic dike studied in thin section is shown by a cross. Each cross is accompanied by letters and numerals. The upper capital letter indicates the grade of metamorphism: *M* = metadiabase, *G* = green schist facies, *AE* = albite-epidote amphibolite facies, *A* = amphibolite facies; transitional types are shown by horizontal arrow. The intermediate numeral indicates the mineral assemblage in the dike and corresponds to columns in Table 1. The lowest letter and number indicates specimen number; those preceded by *W* were collected by White, those preceded by *B* were collected by Billings.

feet thick. Plutonic rocks, which are not abundant, have been mapped as the French Pond granite and the Ryegate granodiorite (Fig. 1). Three major faults divide the area into four tectonic units; these faults from east to west are the Northey Hill, Ammonoosuc, and Monroe (Fig. 1). The rocks in the various blocks are intensely folded. The orogeny and the metamorphosed mafic dikes are probably late Devonian.

A belt of low-grade metamorphism (chlorite zone of the British classification, green-schist facies of Eskola's classification) extends northeasterly from the south-central margin of the quadrangle to the northeast corner. The grade of metamorphism increases both to the east and west; along the west and east margins of the quadrangle the rocks show middle-grade and high-grade metamorphism (staurolite and sillimanite zones of the British classification, amphibolite facies of Eskola's classification.)

West of the belt of metasedimentary rocks in the chlorite zone there is a progressive increase in the grade of metamorphism. The biotite, garnet, and staurolite isograde appear in orderly succession (Fig. 1). East of the belt of rocks in the chlorite zone some of the metamorphic zones have been eliminated along the Ammonoosuc thrust because rocks in the chlorite zone on the west have been thrust over rocks in the staurolite zone on the east. The metasedimentary rocks around the French Pond granite lie in the sillimanite zone.

Petrography of the Mafic Rocks

General Statement.—Metamorphosed mafic dikes are present throughout the Woodsville quadrangle, but are far more abundant in some places than in others. East of the Ammonoosuc thrust they nowhere constitute more than a few per cent of the bedrock. West of this thrust they make up 15 per cent or more of the bedrock in much of the area between the Ammonoosuc and Monroe faults. Areas where they are even more abundant, making up 20 to 50 per cent of the bedrock, are shown in Fig. 1 by a special symbol.

An area of extraordinary dike intrusion in the center of the quadrangle has been distinguished on Fig. 1 as the Leighton Hill dike complex. In this area all the outcrops are metamorphosed mafic rocks, and no exposures of metasediments were encountered. Outcrops a few tens of feet apart generally contain rocks of different structure and texture. Locally contacts between different types may be observed. For this reason the rocks in this area are believed to represent a dike complex rather than a single massive body of metamorphosed igneous rock. The eastern and southern boundaries of the area as mapped are gradational, but the western boundary is abrupt.

The metamorphosed mafic dikes are uncommon west of the Monroe fault. A few were observed within a mile of the fault, but farther west they are rare.

The dikes range in width from a fraction of an inch up to as much as 50 feet. Most of them are parallel to the schistosity of the adjacent country rock. They characteristically crosscut the bedding. Although a few appear to be folded, many of them truncate the fold axes. Some that are parallel to the bedding are technically sills, but to conserve space all the intrusives will hereafter be referred to as dikes.

General lithologic features.—The metamorphosed mafic dikes show an extraordinary diversity in color, structure, texture and mineralogy. They are gray, dark-green, or black. In general, dikes over 10 feet thick are massive, whereas the thinner dikes are schistose. Some thick dikes have massive interiors bordered by schistose margins. In many of the massive dikes relic phenocrysts are preserved; originally plagioclase, augite, or hornblende, these phenocrysts are now plagioclase, hornblende, epidote, or some other secondary mineral. A coarse texture in some rocks, especially in the Leighton Hill dike complex, is considered to be a relic from the original igneous rock. Although the texture of many of the rocks is lepidoblastic or granoblastic, some are ophitic or hypidiomorphic granular.

The mineralogy is also very varied. This is due primarily to differences in the grade of metamorphism and to changes in chemical composition during metamorphism, chiefly variations in the amount of introduced carbon dioxide. Theoretically in an area such as this, where the metasedimentary rocks show progressive metamorphism from the chlorite zone to the sillimanite zone, the metamorphosed mafic rocks should similarly range from green schists through albite-epidote amphibolites to amphibolites (Turner, 1948,p. 76–98). In general, this expectation is fulfilled. Although chemical changes and structural differences introduce numerous complications, the classification of the metamorphosed mafic rocks adopted here is primarily mineralogical. Nevertheless, the detailed classification used in this paper has a genetic basis that is presented in the section dealing with metamorphism. Although many hundreds of dikes were observed in the field, hand specimens were collected for only about 100. From these, 57 thin sections were prepared and studied to form the basis of the paper. The location and lithology of those dikes studied in thin section are shown on Fig. 1. The reader should not be misled into thinking that the dikes are more abundant near the faults. The distribution merely indicates that the dikes were more intensively studied in these areas in order to compare them on opposite sides of the faults.

The metamorphosed mafic rocks may be grouped into four main cate-

TABLE 1. ESTIMATED MODES OF METAMORPHOSED MAFIC DIKES

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
Number of sections	7	9	4	9	2	2	1	1	1	1	2	1	1	1	4	1	2	6	2
Quartz	6	1	10	5	13		5	25	20	35	40		20	30		5	23	2	3
Plagioclase	33	26	28	35	30	11	5	5		25	x	30	25	30	35	25	28	38	45
Actinolite	30†	32	3			79	45	10	33										
Hornblende												30	10	10	42	65	40	48	
Epidote*	15	23	15	1	1	8	x	60		10	3	5	5		14	1	x	4	4
Chlorite	8	15	32	43	27		40		5	10	40	25	25	20	2		1	3	30
Calcite	1	t	10	14		1			17		13		10	10			3	1	12
Ankerite					23														
Biotite	1	1	t	1					23	20			5		1	2	4	t	
Muscovite				1	6				2						t				
Sphene	t	1	t	t	x		x	x	x				x		1		x	1	1
Leucoxene	1	1	2	t	x	1	2	x			2	10			1			t	2
Apatite	x	t	t					x	x			x	x	x		x	x		
Rutile	t			t															
Magnetite		t	t	t	x			x		x			x	x	3	2	x	1	
Ilmenite	3	x	t	t	x		3	x							1		x	1	1
Pyrite	1	t		t			x				x	x		x	t			1	
Garnet																		1	
Allanite									x										
Sericite	1																		2
Per cent of anorthite in plagioclase	27 (12–35)	9 (5–12)	7 (0–10)	3 (0–10)	5 (0–10)	2 (0–3)	2	10	0	0	?	0	0	16	4 (0–10)	0	33 (15–52)	35 (30–65)	10

* Includes clinozoisite.
† Includes some hornblende.
x—Present in 50 per cent or more of thin sections.
t—Present in less than 50 per cent of thin sections.
Blank—Not seen in any of thin sections.

1. Metadiabase.
2. Albite-actinolite-epidote-chlorite rock.
3. Albite-chlorite-epidote-calcite-actinolite schist.
4. Albite-chlorite-calcite schist.
5. Albite-chlorite-ankerite schist.
6. Actinolite schist.
7. Actinolite-chlorite schist.
8. Epidote-quartz rock.
9. Actinolite-biotite-calcite schist.
10. Quartz-albite-biotite schist.
11. Quartz-chlorite-calcite schist.
12. Albite-chlorite-hornblende schist; transition from green schist facies to albite-epidote amphibolite facies.
13. Chlorite-quartz-albite-hornblende schist; transition from green schist facies to albite-epidote amphibolite facies.
14. Quartz-andesine-chlorite-calcite-hornblende schist; transition from green schist facies to amphibolite facies.
15. Albite-epidote amphibolite.
16. Albite amphibolite.
17. Quartz amphibolite in western part of quadrangle.
18. Amphibolite in eastern part of quadrangle.
19. Albite-chlorite-calcite rock, retrogressed from amphibolite, directly east of Ammonoosuc thrust.

gories, although various transitional types occur: (1) metadiabases; (2) green schist facies; (3) albite-epidote amphibolite facies; and (4) amphibolite facies.

Metadiabase.—The metadiabase might equally well be called metadolerite or metagabbro. These rocks are confined to a small area in the southeast corner of the quadrangle between the Ammonoosuc thrust and the Connecticut River, although elsewhere, notably in the Leighton Hill dike complex, some of the rocks locally retain textural relics suggestive of metadiabase. The metadiabases are gray, dark-green, or yellowish green. Although the texture of some is ophitic, more commonly it is hypidiomorphic granular. The grains generally range from 0.5 to 2 mm. in maximum dimension, but plagioclase phenocrysts 4 mm. long are common and some are 10 mm. long. In the hypidiomorphic rocks the dark minerals tend to be euhedral to subhedral whereas the plagioclase tends to be anhedral, molding around the dark minerals. Such rocks are difficult to distinguish in the field from the massive amphibolites and massive albite-epidote amphibolites. In fact, the general mineralogy of the metadiabases is not unlike that of the amphibolites and albite-epidote amphibolites; only microscopic study distinguishes the metadiabases from the albite-epidote amphibolites.

An average mode of seven thin sections of metadiabase is given in Table 1, column 1. The essential minerals are calcic oligoclase or sodic andesine, amphibole, epidote or clinozoisite, and chlorite. The various minerals are intergrown and overlap in a complex way. Although in most specimens the plagioclase is unzoned, in some it shows normal zoning with the cores more calcic than the peripheral shell.

Two kinds of amphibole are present. One has the following optical properties: biaxial negative, $2V=65° \pm 5°$; $Y=b$, $Z \wedge c=18°$; X = light-yellow, Y = light-yellowish-green, Z = blue-green; $X<Y<Z$; $\alpha=1.660$ (1.655–1.666), $\beta=1.672$ (1.667–1.679), $\gamma=1.680$ (1.673–1.686). The optical properties are so similar to an analyzed common hornblende from an intrusive amphibolite in the Littleton-Moosilauke area (Billings, 1937, p. 513) that this amphibole is considered to be common hornblende with an average alumina content of about 10 per cent. In other specimens the amphibole is colorless or only slightly colored in thin section; it is fibrous in some specimens but massive in others. The optical properties are: biaxial negative, $2V=60°$, $Y=b$, $Z \wedge c=18°$; X = colorless, Y = very light yellowish-green, Z = light-blue; $X<Y<Z$; $\alpha=1.634$ (1.632–1.636), $\beta=1.647$ (1.646–1.648), and $\gamma=1.656$ (1.655–1.657). These data suggest an amphibole that is essentially actinolite, but containing several per cent of alumina. Perhaps it should be called actinolitic hornblende. But for brevity it will be called actinolite in accordance with the usage

followed by Turner (1948, p. 94). Some geologists, such as Wiseman (1934), prefer to call it hornblende. The curves prepared by Foslie (1945) suggest that the ratio $MgO/2Fe_2O_3+FeO+MnO+MgO$ (in molecular per cents) averages 0.67, but ranges from 0.70 to 0.64. The same ratio in an amphibole with similar indices and analyzed by Wiseman (1934, p. 368) is 0.57; the alumina content is 2.35.

Green schist facies.—Rocks with mineralogical assemblages characteristic of the green schist facies occupy a belt three to six miles wide between the Ammonoosuc thrust and a line half a mile west of the Monroe fault. The western mile of this belt, however, also contains some dikes with mineral assemblages indicative of the albite-epidote amphibolite facies, which is discussed on a later page.

The principal rocks in the green schist facies have been assigned to four categories: (1) albite-actinolite-epidote-chlorite rocks, (2) albite-chlorite-epidote-calcite-actinolite schists, (3) albite-chlorite-calcite schists, and (4) albite-chlorite-ankerite schists. Average modes are given in Table 1, columns 2, 3, 4, and 5. It will be shown later that this is a single isophysical series of rocks of differing chemical composition. Initially all the rocks had essentially the same chemical composition, but it was subsequently changed by the introduction of varying amounts of carbon dioxide.

The albite-actinolite-epidote-chlorite rocks are gray, dark-gray, or dark-green rocks. They are commonly massive, rarely schistose; of the nine specimens listed in Table 1, column 1, 7 are massive, 1 is slightly schistose, and 1 is schistose. The plagioclase is albite-oligoclase. The amphibole has the following optical properties: X = colorless, Y = colorless or very light green, and Z = colorless or very light blue-green; $\alpha = 1.629$ (1.619–1.635) $\beta = 1.642$ (1.632–1.648), and $\gamma = 1.654$ (1.645–1.661). It is considered to be actinolite with a small amount of alumina. The curves prepared by Foslie (1945) suggest that the ratio $MgO/Fe_2O_3+FeO+MnO+MgO$ in molecular per cents averages 0.67, but ranges from 0.80 to 0.60.

A problem in nomenclature is introduced by these rocks. They are identical with the actinolitic greenstones of Vogt (Barth, Correns, and Eskola, 1939, p. 356) and assemblage #7 of Turner (1948, p. 94). As shown by the average mode the most abundant mineral is amphibole and in this sense they are amphibolites. However, in all other respects they are unlike true amphibolites, inasmuch as they have considerable epidote and chlorite, the plagioclase is highly sodic, and the amphibole is actinolite. Moreover, because of the facies classification, the term amphibolite is generally restricted to rocks composed chiefly of intermediate plagioclase and common hornblende, relatively rich in alumina.

The albite-chlorite-epidote-calcite-actinolite and albite-chlorite-calcite schists are greenish-gray or dark-green. Of 13 specimens listed in Table 1, columns 3 and 4, all but one are schistose.

The albite-chlorite-ankerite schists are green to dark-green schistose rocks with ankerite porphyroblasts, 0.5 to 1.5 mm. in diameter, which are pink to light-brown where fresh but weather to chocolate-brown limonite. They are extensively developed in the Littleton-Moosilauke area (Billings, 1937, p. 512).

Modes of some uncommon rocks found in the green schist facies of the Woodsville quadrangle are given in Table 1, columns 6, 7, 8, 9, 10, and 11.

Albite-epidote amphibolite facies.—These rocks are confined to a belt one to two miles wide in the vicinity of the Monroe fault (Fig. 1). They are dark-yellow-green, dark-gray, and black; some are massive, some are schistose, and some even preserve a relic ophitic texture. The essential minerals are albite, amphibole, and epidote; the average mode of 4 specimens is given in Table 1, column 15. The optical data for the amphibole in one of these specimens are: biaxial negative, $2V=65°$; $X=$light-yellow, $Y=$brownish-green, $Z=$blue-green; $X<Y<Z$; $\alpha=1.651$, $\beta=1.664$, and $\gamma=1.673$; $Z\wedge c=19°$. These data suggest a common hornblende similar to that analyzed in the Littleton-Moosilauke area (Billings, 1937, p. 513), in which the ratio $MgO/2Fe_2O_3+FeO+MnO+MgO$ (in molecular proportions) is 0.55. This is also in close agreement with the curve given by Foslie (1945) for hornblende in the albite-epidote amphibolite facies. Foslie believes that in these hornblendes the alumina substitutes in part for silica and in part for the magnesium-iron-manganese group in the space lattice.

A few specimens, from the vicinty of the Monroe fault, apparently represent a transition from the green schist facies to albite-epidote amphibolite. Representative modes are given in Table 1, columns 12 and 13. Like the rocks in the green schist facies, they have considerable chlorite and albite. But the amphibole is apparently common hornblende, because the optical properties are: $X=$light-yellow, $Y=$light-yellow-green, and $Z=$bright blue-green; $\alpha=1.663$, $\beta=1.674$, and $\gamma=1.684$.

Amphibolite facies.—The amphibolites have been most extensively studied in the area east of the Ammonoosuc thrust. Here they are massive or weakly schistose, speckled black and white rocks which in some outcrops have plagioclase phenocrysts 1 to 5 mm. long. In some specimens the groundmass is so dense that the individual minerals can be recognized only by the aid of a hand lens, in others the hornblende occurs as black needles 1 to 7 mm. long, surrounded by small white grains of feldspar 1 to 4 mm. in diameter.

Microscopic study shows that essential minerals are intermediate plagioclase (andesine or labradorite) and common hornblende; an average mode is given in Table 1, column 18. In the non-porphyritic specimens the plagioclase is uniform within a single hand specimen, but ranges from An_{30} to An_{42} in different specimens. The hornblende is in needles flattened parallel to the front pinacoid, encloses other minerals poikilitically, and has the following optical properties: biaxial negative, $2V=60°$; $X=$light-yellow, $Y=$brownish-green, and $Z=$blue-green; $X<Y<Z$; $\alpha=1.655$ (1.649–1.660), $\beta=1.666$ (1.659–1.671), and $\gamma=1.675$ (1.669–1.680); dispersion, low; $Z\wedge c=20°$. These data are very similar to those obtained from common hornblende in the Littleton-Moosilauke area (Billings, 1937, p. 513 and 556).

West of the Ammonoosuc thrust the amphibolites are confined to that part of the quadrangle west of the garnet isograd (Fig. 1). An average mode is given in Table 1, column 17. The amphibole is apparently common hornblende, as the indices are $\alpha=1.655$, $\beta=1.665$, and $\gamma=1.675$.

One specimen (Table 1, column 14) is considered to be a transitional variety from the green schist facies to the amphibolite facies. It is rich in chlorite, indicative of the green schist facies, but the plagioclase is fairly calcic and the amphibole is common hornblende; the indices are: $\alpha=1.654$, $\beta=1.666$, $\gamma=1.676$. The hornblende is younger than the chlorite, for the former cuts across the schistosity defined by the latter.

METAMORPHISM OF MAFIC DIKES

General Statement.—In the description of the mafic dikes it was pointed out that many rocks, now mineralogically and chemically different, were originally diabase. The diabase is now represented by at least 7 different mineralogical assemblages; moreover, these rocks possess numerous textural and structural differences. It is the purpose of this section of the paper to inquire into the processes whereby originally similar rocks became so diversified. The three principal factors causing this diversification were: (1) differences in physical conditions during metamorphism, indicated by the progressive metamorphism; (2) changes in chemical composition, chiefly due to the introduction of carbon dioxide; and (3) differences in structural behavior and history.

Metadiabases.—The average mode of the metadiabases of the Woodsville quadrangle is repeated in Table 2, column 1. The essential minerals are calcic oligoclase, amphibole, epidote, and chlorite. The amphibole is both hornblende and actinolite. For comparison, the average mode of 24 metadiabases from the White Mountain magma series in the Mt. Washington quadrangle (Fowler-Billings, 1944, p. 1263) is given in Table 2, column 2; the essential minerals are labradorite and hornblende. The hornblende is secondary after pyroxene, probably due to late magmatic

TABLE 2. MODES OF METAMORPHOSED DIABASE

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Quartz	6	1		1	3	11	10	16	5	16	13				2
Plagioclase	33	55	32	26	35	31	28	30	35	29	30	32	35	39	38
Actinolite	30†		28	32	27	10	3								
Hornblende		20										58	42	61	48
Chlorite	8	6	18	15	12	30	32	37	43	33	27		2		3
Epidote*	15		22	23	14	9	15		1		1	10	14		4
Calcite	1	4			3	9	10	16	14						1
Ankerite										18	23				
Biotite	1	3		1	2				1				1		
Muscovite								1	1	4	6				
Pyroxene		3													
Leucoxene	1	1					2						1		
Magnetite		2											3		1
Ilmenite	3	4			1								1		1
Pyrite	1	1													1
Sphene				1	1								1		1
Sericite	1				2										
Per cent of anorthite in plagioclase	27	57	10	9	5	10	7	10	3	10	5	10	4	26	35

* Includes clinozoisite.

† Includes some hornblende.

1. Average mode of 7 metadiabases, Woodsville quadrangle.
2. Average mode of 24 metadiabases of White Mountain magma series, Mt. Washington quadrangle (K. Fowler-Billings, 1944, p. 1263).
3. Calculated mode of Daly's average diabase metamorphosed to green schist facies without introduction of carbon dioxide.
4. Average mode of 9 albite-actinolite-epidote-chlorite rocks, Woodsville quadrangle.
5. Average mode of Moulton diorite, Littleton-Moosilauke area (Billings, 1937, p. 503).
6. Calculated mode of Daly's average diabase metamorphosed to green schist facies with introduction of 4.4 weight per cent of carbon dioxide, mode recalculated to 100 per cent.
7. Average mode of 4 albite-chlorite-epidote-calcite-actinolite schists, Woodsville quadrangle.
8. Calculated mode of Daly's average diabase metamorphosed to green schist facies with addition of 7.25 weight per cent of carbon dioxide; mode recalculated to 100 per cent.
9. Average mode of 9 albite-chlorite-calcite schists, Woodsville quadrangle.
10. Calculated mode of Daly's average diabase metamorphosed to green schist facies with introduction of 8.8 weight per cent of carbon dioxide; mode recalculated to 100 per cent.
11. Average mode of 2 albite-chlorite-ankerite schists, Woodsville quadrangle.

hydrothermal alterations (Fowler-Billings, 1944, p. 1271). The metadiabases of the Woodsville quadrangle are more extensively altered. The original plagioclase, presumably labradorite, has gone to calcic oligoclase, thus releasing considerable anorthite to form epidote. Much of the hornblende has gone to actinolite. The metadiabases of the Woodsville quadrangle are well on the way to a mineralogical assemblage characteristic of the green schist facies, but have not fully acquired the mineralogical characteristics of that facies. These changes must have taken place during the regional metamorphism.

The restriction of the metadiabases to a relatively small area demands explanation. The adjacent metasedimentary rocks in this area are hornfelses, apparently due to the abundance of the mafic dikes. The resulting rigid block resisted deformation more than the sedimentary rocks in other parts of the quadrangle. Possibly the temperatures were also lower here during the metamorphism than elsewhere in the area. If rigidity of the rocks were the sole explanation, one might expect the metadiabases to be even more abundant in the Leighton Hill dike complex. Some of the mafic dikes in that complex do preserve relics of ophitic and hypidiomorphic granular textures. But the isograds suggest that the temperature during metamorphism was somewhat higher in the region occupied by the Leighton Hill dike complex and thus conditions were favorable for more complete metamorphism.

Green schist facies.—The writers have calculated the expected mineralogy of an average diabase (Daly, 1933, p. 18) when metamorphosed under various conditions. The results are given in Table 2, columns 3, 6, 8, 10, 12, and 14. For these calculations TiO_2 and H_2O were neglected and Fe_2O_3 was converted to FeO. Water was also neglected, on the assumption that it was always available in sufficient quantity to form whatever minerals were demanded by the physical conditions. No attempt was made to calculate the amount of such minor minerals such as sphene, ilmenite, magnetite, or pyrite. It is obvious that the calculated modes will not agree rigorously with the actual modes found in the rocks in the Woodsville quadrangle. The chemical composition of the original mafic rocks of this area must have differed somewhat from Daly's average diabase. Moreover, the chemical composition of the actual minerals must differ to some extent from those used in the calculations. The best that

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12. Calculated mode of Daly's average diabase metamorphosed to albite-epidote amphibolite.
13. Average mode of 4 albite-epidote amphibolites, Woodsville quadrangle.
14. Calculated mode of Daly's average diabase metamorphosed to amphibolite.
15. Average mode of 6 amphibolites from eastern part of Woodsville quadrangle.

can be expected is that there will be a general similarity between the calculated modes and the actual modes.

Table 2, column 3 gives the calculated mineralogical composition of Daly's average diabase metamorphosed in the green schist facies without introduction of carbon dioxide. Table 2, column 4 lists the average mode of 9 albite-actinolite-epidote-chlorite rocks from the Woodsville quadrangle. The calculated and actual modes compare very favorably. Table 2, column 5 lists the average mode of the Moulton diorite of the Littleton-Moosilauke area (Billings, 1937, p. 503); this rock has been metamorphosed so that it now consists of a mineralogical assemblage typical of green schists. It likewise compares favorably with the calculated mode in column 3.

When carbon dioxide is introduced into the rocks, the lime is used to form calcite. With a progressive increase in the amount of carbon dioxide introduced there will be a corresponding decrease in epidote and actinolite accompanied by an increase in chlorite, calcite, and quartz. Table 2, column 6 gives the calculated mode if 100 molecular proportions (4.4 weight per cent) of carbon dioxide are added to Daly's average diabase and the resulting mode recalculated to 100 per cent. Table 2, column 7 lists the average mode of 4 albite-chlorite-epidote-calcite-actinolite schists from the Woodsville quadrangle. The similarity between the actual and calculated modes is striking.

So much carbon dioxide may be introduced that all the lime—except that necessary to form albite-oligoclase—is used up to form calcite. In this case the mode would be that given in Table 2, column 8; 165 molecular proportions (7.25 weight per cent) of carbon dioxide were added. Epidote and actinolite are no longer present, but chlorite, calcite, and quartz have increased. Table 2 column 9 lists the average mode of 9 albite-chlorite-calcite schists from the Woodsville quadrangle. The calculated and actual modes compare favorably, except perhaps for quartz.

If even a greater quantity of carbon dioxide is introduced, it combines with some of the ferrous oxide and magnesium oxide as well as lime to form ankeritic calcite or ankerite. Table 2, column 10 gives the calculated mode if 200 molecular proportions (8.8 weight per cent) of carbon dioxide are added to Daly's average diabase. Table 2, column 11, gives the average mode of 2 albite-chlorite-ankerite schists from the Woodsville quadrangle. The calculated and actual modes compare favorably.

The assumption has been made in the preceding paragraph that ferrous and magnesium oxides would not combine with carbon dioxide until all the available lime—except that necessary for albite-oligoclase—has been used. Actually, of course, the relations may be more complex and carbon dioxide may begin to combine with ferrous and magnesium oxide at an earlier stage.

We conclude that under physical conditions characteristic of the green schist facies 4 different mineralogical assemblages have been produced from original diabases due to variations in the amount of carbon dioxide introduced. The mineralogical and chemical changes, however, are not the only factors involved. The metamorphosed mafic dikes in the green schist facies range from massive to schistose. The massive dikes locally preserve relics of the original igneous textures. In general, the massive dikes are albite-actinolite-epidote-chlorite rocks without carbonate, whereas the schistose dikes are rich in chlorite and carbonate. Considerable carbon dioxide was added to the schistose dikes. Moreover, several dikes carefully studied in the field have a massive interior with little or no carbonate but a schistose border rich in carbonate. There is clearly a problem to determine which was cause and which was effect. Did the presence of schistosity permit the introduction of carbon dioxide or did the introduction of the carbon dioxide and the consequent formation of carbonate and chlorite permit the development of schistosity?

In general, therefore, small dikes were metamorphosed into albite-chlorite-calcite schists. Large dikes, on the other hand, developed into rocks devoid of carbonates, although the margins may have become an albite-chlorite-carbonate schist. Size, however, is not the only factor. In some areas dikes of equal size have suffered very different fates, one forming massive albite-actinolite-epidote-chlorite rocks, the other developing into an albite-chlorite-calcite rock. Possibly the orientation of the dike relative to tectonic forces may have played a role; one dike would be sheared whereas its neighbor would not. This phase of the problem needs further investigation in the field.

The 31 modes averaged to give columns 1, 2, 3, 4, and 5, Table 1, represent diabase dikes that are essentially the same in chemical composition as the original igneous rock, except for the addition of water and in some instances carbon dioxide. On the other hand, the 8 modes given in Table 1, columns 6, 7, 8, 9, 10, and 11 cannot have been diabase originally, unless even more extensive changes in chemical composition took place. The rock represented by the mode given in column 10 was presumably originally a quartz diabase. Columns 6 and 7 may represent highly mafic portions of an original diabase dike. But the four modes given in columns 8, 9, and 11 are rich in quartz and mafic minerals, whereas they have little or no plagioclase; soda is very low in these rocks. If they were diabase prior to metamorphism, extensive changes in chemical composition have taken place and most of the soda was removed.

Some of them may have been sedimentary rocks that were misinterpreted to be dikes in the field. Unfortunately, the writers have not had an opportunity to restudy these problems in the field, but future investigators in this and similar areas should appreciate the several possibilities.

Albite-epidote amphibolite facies.—The calculated mineralogical composition of Daly's average diabase metamorphosed to albite-epidote amphibolite is given in Table 2, column 12; the average mode of 4 albite-epidote amphibolites from the Woodsville quadrangle is given in Table 2, column 13.

In addition to the four dikes that were studied microscopically and classified as albite-epidote amphibolite, three additional dikes (Table 1, columns 12 and 13) are considered to be transitional from the green schist facies to the albite-epidote amphibolite facies. As shown on Fig. 1, all seven of these rocks lie within a narrow belt about one mile wide near the Monroe fault. The proximity to the fault is believed to be a coincidence and to have no genetic significance. Rocks characteristic of the green schist facies lie within this same belt (Fig. 1). In the southern part of the quadrangle this belt lies east of the biotite isograd, but in the northern part it lies to the west. According to Turner (1948, p. 88) the albite-epidote amphibolites coincide approximately with the garnet (almandite) zone, that is they lie between the garnet and staurolite isograds. In the Woodsville quadrangle they lie in the higher grade part of the chlorite zone and in the biotite zone.

Amphibolite facies.—The calculated mineralogical composition of Daly's average diabase metamorphosed to amphibolite is given in Table 2, column 14, and may be compared with the average mode of 6 amphibolites from the eastern part of the quadrangle given in Table 2, column 15.

Unfortunately, the precise data are lacking in this area on the spatial relations of amphibolites to the isograds defined by the argillaceous sedimentary rocks. In the eastern part of the quadrangle the displacement along the Ammonoosuc thrust has brought the green schist facies into direct contact with the amphibolite facies. In the western part of the quadrangle not enough thin sections were studied to define precisely the eastern limit of the amphibolites. The scanty data available are merely suggestive. At the garnet isograd (Fig. 1) a rock belonging to the green schist facies is partially transformed into an amphibolite (Table 1, column 14). An amphibolite—actually a quartz amphibolite with garnet—lies one-third of a mile east of the staurolite isograd. These data suggest that the amphibolites may develop within the garnet zone, but the data are admittedly inconclusive. Turner (1948, p. 76), however, likewise places the first appearance of amphibolites within the garnet zone.

Retrograde effects.—Rather striking retrograde effects are found in the metasedimentary rocks in a belt 500 to 1000 feet wide directly east of the Ammonoosuc thrust (White and Billings, 1950). Amphibolite dikes in this zone have retrogressed to albite-chlorite-calcite schists. Two such dikes, specimens B 6 and B 60, are indicated in Fig. 1. An average mode is given in Table 1, column 19.

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A SIMPLE TECHNIQUE FOR THE STUDY OF THE ELASTICITY OF CRYSTALS*

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ABSTRACT

A brief review is given of the use of the composite piezoelectric oscillator for determination of the elastic constants of small samples. A few preliminary results for single crystals have been obtained, for comparison with older measurements.

INTRODUCTION

Of some hundreds of important crystals, the elastic constants are known for about forty, including artificially-grown metals and alkali halides (Birch, et al., 1942, pp. 66–70). Many of the measurements on natural crystals of low symmetry were made by Voigt in an arduous study of the static bending and twisting of crystalline bars or plates. Voigt probably accomplished nearly all that is possible with natural crystals by these methods, for which fairly large samples are required. Recent advances have been possible partly because of the production of relatively large synthetic crystals, partly because of the development of new methods which permit the use of smaller samples. It is the purpose of this note to draw attention to the existence of a method applicable to many minerals which rarely occur in large sizes. Mineralogists, who have first access to mineral specimens and are familiar with problems of crystal orientation and crystal symmetry, may be encouraged by the simplicity of the experimental arrangements to undertake further studies in this relatively neglected field.

In recent years, a number of dynamic methods of measuring elasticity have come into use; the one to be described is the method of the composite piezoelectric oscillator which has been employed in a number of important investigations by Quimby (1925), Balamuth (1934) and others (Cady 1946, p. 484). In this technique, a small sample in the form of a prism of uniform cross-section, is cemented to a piezoelectric crystal, usually quartz, whose natural resonant frequencies are known. Under suitable restrictions, determination of the resonant frequency of the composite oscillator then leads to a value for the resonant frequency of the sample alone, from which, together with the dimensions and the density, one elastic constant or one combination of elastic constants, may be computed.

There is no way of avoiding the inherent complexity of crystal elas-

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ticity; in order to determine all of the elastic constants, there must be at least as many measurements, for rods of different orientations and for vibrations in different modes, as there are independent elastic constants. The other remaining difficulties are those of cutting and orienting the sample bars. In preliminary studies, samples as short as about 7 mm. have been used, but there seems to be no reason to take this as a lower limit. Samples several millimeters long should be obtainable from a considerable number of hitherto unstudied crystalline varieties.

THEORY

A theory of two- and three-part composite oscillators may be found in the papers by Balamuth (1934) and Rose (1936). Consider two rods of the same, uniform, cross-section, and let rod 1 be the sample, rod 2 the piezoelectric crystal. Let the masses of the two rods be m_1 and m_2, respectively, their individual frequencies, f_1 and f_2. The frequency f of the composite oscillator formed by cementing these rods end to end is given by (Rose, 1936):

$$(1)\quad m_1 f_1 \tan \pi f/f_1 + m_2 f_2 \tan \pi f/f_2 = 0.$$

Thus if f and f_2 can be determined experimentally, f_1 may be found by (1). This relation is much simplified if f, f_1 and f_2 are all nearly equal. If f_1 and f_2 are within about 10% of f, then the following equation is valid to within a per cent or so:

$$(2)\quad f_1 = f + (f - f_2) m_2/m_1.$$

This convenient approximation to (1) is often sufficiently exact.

The piezoelectric crystal may be cut so that it will be excited in extensional vibrations or in torsional vibrations. For extensional vibrations of thin bars, the lowest frequency of the sample alone, f_1, is related to its length l and density ρ by the relation, $2f_1 l = \sqrt{E/\rho}$, where E is the "Young's modulus" in the direction of the axis of the rod. For bars having a length less than three or four times the diameter, a correction is required, which has been worked out for isotropic materials by Bancroft (1941). Assuming that this theory is approximately valid for crystals, we may conclude that the correction is less than 1% so long as the length is greater than twice the diameter, and "Poisson's ratio" is less than 0.3. There is no correction of this kind for torsional vibrations of rods of circular cross-section, for which the lowest frequency is related to the modulus of rigidity or torsion G about the axis of the rod according to $2fl = \sqrt{G/\rho}$. Equations (1) or (2) will hold approximately even if the cross-sections of sample and piezoelectric crystals are not exactly equal, except that for torsional vibrations the moments of inertia of the sections about the axis should be substituted for the masses. The shape of the cross-section is not of consequence for thin bars in extensional vibration, but the theory of torsional vibration becomes more complicated for sections of other than circular shape (Cady, p. 113). Consequently the most useful shape for samples is the circular cylinder, preferably having its length several times its diameter.

Equations (1) and (2) are not restricted to the lowest or fundamental modes, and it is often more convenient to excite the driving crystal in an overtone than to employ a smaller crystal for which the same frequency would be the fundamental; on the other hand a large ratio of masses, m_2/m_1, exaggerates the effect of the frequency difference $f-f_2$.

The relations between E and G, the orientation and the individual elastic constants of the sample are given by Voigt, Cady, and for certain cases, by Wooster.

EXPERIMENTAL ARRANGEMENTS

The detection of the resonant vibration of a piezoelectric crystal depends upon the fact that at resonance, the impedance of the crystal undergoes a marked reduction which may be recognized by a variety of methods. The resonance curves for quartz and for most single crystals are extremely narrow, so that the first requirement is a source of variable frequency with fine frequency control through the correct range. The oscillator should supply a voltage of the order of 10 or more volts across the crystal and the series resistor R, of perhaps 0.5 megohm (Fig. 1).

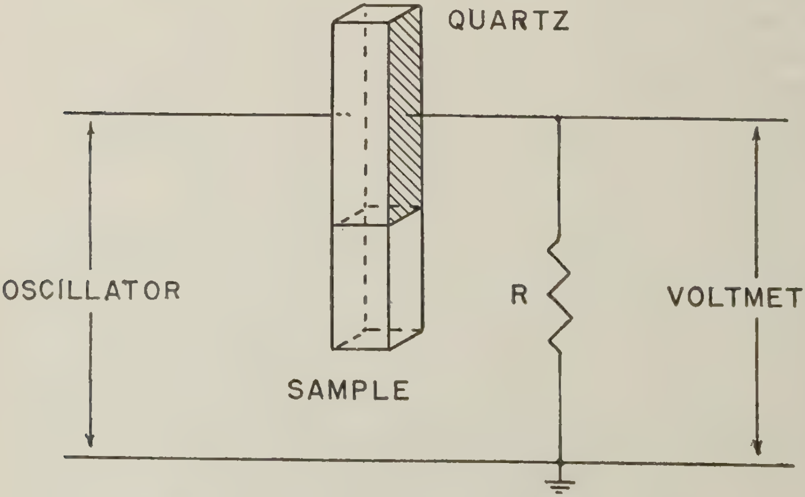

FIG. 1. Schematic circuit for detection of resonance of the composite oscillator.

A vacuum-tube voltmeter, with a short-period meter, or a cathode-ray oscilloscope, is connected across R. As the frequency of the voltage is varied, resonance is indicated by a sharp swing of the meter or a rapid modulation of the pattern on the oscilloscope screen. The voltage across R, small for frequencies away from resonance, may rise as much as several volts at resonance.

There are various methods of calibrating the frequency scale, which depend upon the available equipment. A number of standard frequencies are broadcast from the Bureau of Standards. Crystals which resonate at accurately-determined frequencies may be obtained from a number of manufacturers. Unless the dimensions are unusually well-determined, it is not worth while to measure the frequency more accurately than to about 1/1000; other uncertainties are likely to approach 1%. In the present work, the 1000-cycle standard frequency signal of the Cruft Laboratory is used as the reference frequency, and applied to the horizontal axis of an oscilloscope. The signal from the variable frequency source is applied to the vertical axis. Observation of the Lissajous figures produced by this arrangement gives a great number of fixed points on the frequency scale, at 1000-, 500-, or even 100-cycle intervals. A precision capacitor is used for interpolation between these fixed frequencies.

For the minimum disturbance of the resonant frequency, the oscillator should be supported at displacement modes. In the fundamental extensional mode, the plane normal to the axis through the center of the rod is a node for longitudinal displacement, but the lateral displacement is a maximum in this plane. The lateral motion is small for quartz, however, and support by a spring clip at this position has a negligibly small effect on the frequency. It is not even essential to place the clip at the exact center, though the amplitude of resonance is reduced if the departure from the central position is marked. The two leaves of the clip are separately fastened to an insulating strip, and are used as the electrical connections to the quartz oscillator.

For extensional vibrations, X-cut quartz bars of square cross-section are convenient. The faces normal to the electric axis are coated with a thin conducting film: sputtered gold surfaces are most durable, but serviceable electrodes may be made either by pasting on thin metal foil or by painting with a conducting paint such as silver paste or even Aquadag. Rose (1936) has described a quartz oscillator of circular cross section which may be excited in torsional modes. This oscillator has four electrodes, connected in pairs, for which foil is well adapted. The crystal may be supported by clips at its central plane without disturbance of the frequency.

The composite oscillator is formed by cementing the sample of unknown frequency to a quartz oscillator. A number of cements will serve, such as DeKhotinsky wax or shellac flakes, and probably many others. The shellac is softened on an electric plate, and applied sparingly to the two ends to be joined, which are also warmed. The two rods are then pressed together and allowed to cool. A strong joint is not required, since, if the frequencies are properly matched, the interface is a plane of zero stress, but there should be a minimum of space between the rods, and a minimum of foreign material. Soft waxes or cements having high internal friction are undesirable. The composite oscillator is placed in the spring clip with the quartz oscillator centered on the clip (for the fundamental modes), just as for the quartz alone.

RESULTS

No systematic investigation has as yet been completed, but the method has been tested upon a set of samples of single crystals originally selected by Professor Charles Palache for use in investigations of compressibility by Professor P. W. Bridgman, in whose papers (1925, 1928, 1949) brief descriptions may be found. These samples ranged in length from about 1 inch down to 0.5 inch; some were ground to a uniform cross-section, others were bounded by natural prism faces. A few were too irregular for the present purpose. In no case were there enough samples for the determination of all of the elastic constants, but comparison with measurements by other observers is possible in a number of instances.

A set of X-cut quartz oscillators of square cross-section was prepared, ranging in length from 1 inch to 0.275 inch, in diameter from about 0.2 inch to 0.125 inch, and in frequency from 100 KC to 370 KC. Successive oscillators differed in length and frequency by about 10%. The faces normal to the X-axis were given a light coating of Aquadag.

Each sample was coupled to different quartz oscillators until a composite oscillator was found whose frequency fell between the frequencies of two adjacent quartz oscillators. Equation (2) then becomes a formula for interpolation, and gives the frequency of the sample with an error usually less than 1%. Greater precision could be obtained by more careful matching of frequencies and of cross-sections.

As an example of the calculation, consider the sample of beryl with axis normal to "*c*." Its mass was 1.205 gm., length 0.713 inch. The composite oscillator, formed by joining this with a quartz oscillator of mass 0.515 gm., 0.450 inch long, with a natural frequency of 234,670 cps, had its resonance at 244,870 cps. Equation (2) then gives 249,240 for the frequency of the beryl alone. The frequency of the composite oscillator formed by joining the beryl with a quartz oscillator of mass 0.415 gm. and frequency 262,630 cps was 253,000 cps. This gives 249,690 cps for the frequency of the beryl alone, about $\frac{1}{2}\%$ different from the value found with the other oscillator. The mean value, 249,460, gives the velocity, $2fl = 9.03$ km/sec.

The measurements are summarized in Table 1. The velocity which is tabulated is $2\,fl$, with no correction for finite diameter, but the rato of length to diameter was always greater than three or four. The modulus is obtained from $2fl = \sqrt{E/\rho}$. The compliance, given in the last column, is the reciprocal of the modulus. All of these results are for extensional vibrations.

TABLE 1. ELASTIC CONSTANTS OF EXTENSIONAL VIBRATIONS DETERMINED FROM FREQUENCIES

Crystal	Density	Velocity Km/sec	Modulus, in 10^{11} dyne/cm²	Compliance, in 10^{-13} cm²/dyne
Beryl				
par. *c*-axis	2.668	9.03	21.8	$s_{33} = 4.60$ (4.71)[1]
perp. *c*-axis	2.683	9.03	21.9	$s_{11} = 4.57$ (4.42)[1]
Tourmaline, par. *c*-axis				
black	3.091	6.88	14.6	$s_{33} = 6.84$ (6.24)[1] (6.0)[2]
yellow	3.028	7.40	16.6	$s_{33} = 6.03$
pink	3.031	7.47	16.9	$s_{33} = 5.91$
Topaz				
a-axis	3.538	8.05	22.9	$s_{11} = 4.36$ (4.43)[1]
b-axis	3.545	9.12	29.5	$s_{22} = 3.38$ (3.53)[1]
c-axis	3.548	8.49	25.6	$s_{33} = 3.91$ (3.84)[1]
Pyrite, *a*-axis	(5.018)[4]	8.68	37.8	$s_{11} = 2.65$ (2.69)[3] (2.89)[1]
2nd sample		8.72	38.2	$s_{11} = 2.62$
Andradite	3.482	8.13	23.0	orientation undetermined
Garnet (pyrope)	4.122	7.57	23.6	
Spodumene	(3.186)[4]			
a-axis		8.96	25.6	
b-axis		7.50	17.9	$s_{22} = 5.6$
Albite-olivine glass, (25% Mg_2SiO_4)	2.51	5.65	8.03	

[1] Voigt (1928).
[2] Osterberg and Cookson (1935).
[3] Birch and Bancroft, unpublished.
[4] Handbook values, after Berman.

Where possible, values obtained by other methods are given in parentheses; most of these are by Voigt, and show about the same amount of discrepancy as has been found by other observers who have used dynamical methods (Cady, Chapter VI). A difference between Voigt's isothermal values and the present adiabatic ones is to be expected, but this is of the order of a few parts in a thousand.

Table 2 shows results for a group of X-cut quartz rods all nearly 1 inch long, and of approximately square cross-section, about 0.15×0.15 inch. The mean value for 2 *fl* is 5.457 km/sec; with $\rho = 2.654$ we obtain $s_{11} = 12.64 \cdot 10^{-13}$ cm²/dyne, in good agreement with the results of other investigators (Cady, 1946, pp. 135–138).

TABLE 2. LOWEST EXTENSIONAL FREQUENCY OF X-CUT QUARTZ BARS

Length inches	Frequency KC	$v = 2fl$ km/sec
1.081	99.31	5.453
1.085	99.00	5.457
1.058	101.53	5.457
1.142	94.35	5.473
1.086	98.91	5.457
1.086	98.82	5.452
1.080	98.89	5.486 (poor)
1.084	98.95	5.449
1.123	95.54	5.450
1.086	99.07	5.466
1.122	95.70	5.455
1.086	98.95	5.458

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THE MAKING OF A MAGMATIST

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Abstract

Professor Larsen's studies of lavas, minor intrusions, and batholiths in the San Juan region of Colorado and in Southern California are presented as the basis of his development of magmatist views. Anti-magmatist views on the formation of batholiths are discussed, and also the magmatist view that batholiths are formed by refusion in a tectogene.

He probably scorns a label, preferring to be regarded as one who keeps an open mind, as indeed he is (Larsen, 1950), but it is no great distortion of the truth to say that Professor Larsen is a magmatist. He would not, at least, regard this designation as a term of contempt, nor should he, for he acquired magmatist views as a result of long years of arduous field studies in difficult but rewarding terranes and of equally arduous investigation of materials and collation of facts gleaned in the field.

The San Juan

In 1909 Larsen joined Whitman Cross in a comprehensive study of the geology of the San Juan region of Colorado where Cross had already been working for fifteen years. During eighteen additional years, with some interruptions occasioned by the first World War, field investigations were continued by Larsen, partly in association with Cross, but with many other associates as well, the most notable of whom was C. S. Ross. Many publications, including seven folios, had been written by Cross, and in 1935 appeared a "brief review" of the geology of the region written by Larsen (Cross and Larsen, 1935).

The San Juan region was found to be one in which volcanic activity, dominantly of Tertiary age, was very prominent, indeed most of the mountains of the region have been carved in great accumulations of lavas and ejectamenta, extruded at intervals throughout a long period of time. The lavas range from basalts through several varieties of andesites, latites and quartz latites to typical rhyolites. There is no simple order of eruption. Several volcanic series are distinguished, each of which includes a wide range of compositions and in some instances they were separated from each other by intervals of erosion.

It would seem impossible that anyone would deny that these lavas represent igneous magmas, and it is to be hoped that anti-magmatists are merely being facetious when they state that the existence of granitic magma is based entirely on assumption. Larsen, at least, and all of his associates, without exception or reservation, concluded that the field evidence was decisively in favor of the formation of these rocks from molten magmas.

Magnificent studies of phase petrology by Professor Larsen and associates (1936, 1937, 1938) have led them to some very definite conclusions as to the courses of crystallization in the various types of lava and in the series of lavas as a whole. These indicate that a prominent control over the derivation of the series is fractional crystallization of liquid magma, but assimilation of foreign material and actual mixing of magmas are also suggested by the observations.

In addition to the lavas of the region there are bodies of rock showing the same range of chemical composition as the effusive lavas but having forms and relations to their surroundings which lead to their description under the accepted terms dikes, sills, sheets, laccoliths, and stocks. These yield varying degrees of evidence of having been insinuated into existing rock masses. Larsen and associates have thus been led to regard them as intrusive rocks of magmatic origin. In the smallest of bodies, as for example in thin sheets or sills, these rocks may differ scarcely at all, in type of crystallization exhibited, from their chemical equivalents among the lavas. Thus some of the sheets are, like the similar lavas, referred to as andesites, but in larger masses the degrees of crystallinity and of granularity may depart notably from those of the lavas and typical granular rocks (diorite among others) are found in some larger stocks and laccoliths. The passage from the one extreme of partly glassy rocks to holocrystalline granular types is so well exhibited that the crystallization of all of the masses from igneous magmas analogous to the lava flows has been accepted by Larsen.

The conclusion seems reasonable, but as soon as we move down more than a foot or two from the surface anti-magmatists begin to balk at the suggestion that rocks in these situations were formed from magmas. It must be supposed therefore that some of them, at least, would regard these hypabyssal minor intrusions as formed by some variety of metasomatic replacement. What evidence they might be able to adduce in favor of a replacement origin of the masses is a question that must remain unanswered here. It is to be noted, however, that at the borders of the masses replacement effects are to be found locally and some valuable ore deposits have been formed in connection with these effects, but that these relations can be taken as an indication of the replacement origin of the "intrusive" itself is a view which Larsen would regard as unconfirmed by the field evidence, as would a great many others who have investigated the ore deposits and their relations and who were intensely interested in problems of metasomatic replacement.

Rather they regard the evidence as favoring the concept that igneous magmas have come from still greater depths in the earth and have formed the hypabyssal intrusives and the lava flows. Speaking of the stocks, Cross and Larsen (1935, p. 126) say that "they have a closer connection with the deeply buried parent magma chamber."

Do we have any clue as to nature of such materials at greater depths?

SOUTHERN CALIFORNIA

Even before he began his work in the San Juan region Professor Larsen, as an assistant to W. C. Mendenhall, had in 1906 studied a part of the batholith of Southern California, and in 1930, the same year in which he made his final corrections to the field studies in the San Juans, he began a systematic study of this batholith. As formerly, he had a number of collaborators, the field investigations extended over several years, and he was able to incorporate in his report the observations of others not directly associated with him in his investigations. If his conclusions are correct, the batholith, though most remote from the San Juan intrusives in both space and time, may nevertheless be regarded as taking up the story of igneous action about where the San Juan intrusives leave off, but perhaps with significant overlap, for there are minor intrusions satellitic to the main batholith, itself a complex of intrusives. These minor intrusions may be likened to the stocks and laccoliths of the San Juans. The main batholithic complex may then be something of the nature of a magma chamber analogous to that which supplied the magmas of the San Juans.

The batholith of Southern California is itself presumed to be derived from a still deeper magma chamber. Larsen (1948, p. 171) says "A magma of relatively uniform gabbro was formed at depth in the core of the ranges. This gabbro was slowly differentiating, having at any time in its history an upper part essentially the same throughout the length of the batholith as to chemical composition and kind and amount of suspended crystals. From time to time diastrophism moved the magma toward the surface. . . . After emplacement of the gabbro, little further movement took place until the upper part of the magma had the composition of a tonalite. Then many local and several widespread movements emplaced the several tonalites. . . . After intrusion of the tonalites no widespread earth movement took place until the upper part of the remaining magma reached the composition of a granodiorite. The Woodson Mountain granodiorite was then injected over much of the length of the mountain range. The small volume of granite in the batholith may result from a scarcity of residual granite liquid or from the fact that during the granitic stage diastrophism was not favorable to movement of the magma toward the surface."

The picture he paints of an order of intrusion related to and controlled by an order of crystallization is supported by a wealth of data on phase petrology. He thus finds that, in determining the relations between the various rock types, the same factor, crystal⇌liquid equilibrium, has been in control.

COMMENT

The above is in skeleton outline the story of the making of a magmatist. Though it spans a period of forty-five years it by no means presents or even suggests all the geological phenomena and regions studied by Larsen. Wide and varied has been his experience, yet it appears to have served, on the whole, only to yield more evidence in favor of magmatism.

It is earnestly to be hoped that no one will make the insulting suggestion that Larsen heard about magmas from his teacher but has never given the subject any real thought himself, a suggestion that has been made of another investigator not less experienced and distinguished than Larsen. Daly (1949) has recounted, in his inimitable manner, the story of how he himself, as a young man, diligently exposed himself to all schools of thought on eruptive rocks, yet turned out a magmatist. To be sure antimagmatists tell a similar story of how they have reached their convictions, but it would be most desirable if they would refrain from adding the suggestion that the magmatist had no field experience or failed to think about what he did see in the field.

The question might be raised as to whether Larsen would have been a magmatist if he had studied the California batholith before he studied the San Juan lavas. Actually he did have some prior experience in the California field, but it was relatively limited, and the question is, of course, one to which there can be no reply, nor can it be profitably discussed.

A more useful question is one that asks whether his conclusions regarding the California batholith will withstand the test of time. To this question only time can give a final answer, but it can be profitably discussed, for it is such discussion, and investigation arising from it, that will enable an answer.

In his conclusions, set forth above, Larsen has shown himself to be not only a magmatist but, as far as the California batholith is concerned at least, a special brand of magmatist, one who regards the evidence as pointing to the parental nature of basaltic magma, a view of which Daly has long been the leading proponent.

We shall discuss the various possibilities by turning first to the view at the other extreme. The batholith of Southern California was, let us say, emplaced by solid diffusion. No liquid or fluid of any kind was involved in its formation. Atoms or ions diffused through crystal lattices and along intergranular boundaries under the influence of some assumed potential or driving force, and built up the mass as we find it. Apart from the convenient assumptions already made that this process would be effective at all, on what basis are the additional assumptions made that it would give associations of crystalline phases (rock types) that so closely imitate the results that would be obtained if the control over these associations were crystal-liquid equilibrium?

There is also the objection based on the existence of zoned crystals in the Californian batholith. A chemical potential exists in such crystals tending to make them of uniform composition, but it has frequently been unable even to blur boundaries between zones which would have required diffusion only through hundredths of a millimeter for their complete merging. Yet the ionic diffusion must take place through distances measured in kilometers in order to feed the assumed growth of the crystals by this method. The attempt to explain the persistence of zoning by assuming a lesser facility of ionic diffusion within the crystal lattice than in the intergranular film must be regarded as unsuccessful. No less an authority on crystal structure than Buerger (1948, p. 114) claims that it requires much more energy to cause an ion to escape from a lattice than to wander in the lattice. *Per contra*, then, an ion assumed to be already in an intergranular film will have a much smaller tendency to wander foot-loose in that film than to get back into the lattice, where it can then wander more freely. Even if this reasoning should prove erroneous, and there really is a greater freedom of wandering in the supposed film, it should be realized that this excess of mobility requires to be enormous, considering the enormous discrepancy between the distances involved, kilometers versus hundredths of a millimeter. This becomes increasingly apparent when it is realized that in diffusion effects it is the square of the distance that counts. No, solid diffusion is not a satisfactory way of getting batholiths with the mineral associations and the mineral textures they exhibit, and it is probably too slow a process to yield a mass of batholithic dimensions in the time that can reasonably be allotted to the formation of these masses, long though it may be.

Nor can it be said that the case for formation of a batholith through the agency of an ichor or mineralizers is significantly stronger. The question of time is probably not so critical here, for the bodily movement of material through the minute openings that may develop in a rock mass undergoing deformation, while undoubtedly slow, must be of an order of magnitude greatly in excess of that characteristic of ion diffusion. However, the same objections that have been raised against ion replacement apply here as well, in the matter of the actual petrographic character of the materials produced. A replacement process of any kind should be completely lacking in any tendency to produce an association of minerals in definite or approximately definite proportions. The feldspathization of a quartzite, for example, should go to completion, except in so far as it was accidentally interrupted. It would not characteristically cease while some quartz still remained, and as for coming to a close when there is established that nice balance between quartz, potash feldspar, and albitic plagioclase characteristic of the usual granite, that could be regarded only as a miracle of replacement (Chayes, 1950). The relative proportions of solid phases have no effect on chemical equilibria, the mass of a solid phase being constant whether it be present as a trace or in dominant amount. On the other hand, the relative proportions of the components of a liquid in equilibrium with solids is very definitely controlled, and granites tend to have a constant mineral composition because they are formed by the consolidation of liquids whose composition has been subject to this crystal-liquid equilibrium control. The liquid so prepared can drench surrounding rocks and thus granitize them, but this is a wholly subsidiary process.

The whole batholithic picture points to formation as a result of the intrusion and consolidation of liquid magma. Granitic magma has been shown with reasonable certainty by laboratory studies to be the normal residual liquid from the fractional crystallization of the polycomponent system of rock-forming oxides, a relation that is confirmed by observations upon rocks. But the very facts that indicate this relation for granitic magma offer equal support to the possibility that it would be the first liquid formed in the remelting of a heterogeneous massif. There are some magmatists who favor this view of the formation of batholiths. There is no doubt that Waters (1948, p. 107) had these considerations in mind when, emphatically rejecting replacement of any kind as the controlling process in the formation of batholiths, he proposes that they are formed by selective fusion of granitic gneisses and sialic sediments forced down to levels of high temperature in a tectogene. A process of this general character has been suggested by others as well. Let us consider the probable results of such a process.

If the temperature of the down-warped sial were raised to a value such that quartz, potash feldspar, and sodic plagioclase, together with such water as might be present, could form a liquid, that liquid would, of course, have these components in granitic proportions, for the control of crystal-liquid equilibrium over the composition of the liquid is just as effective in selective fusion as in fractional crystallization. The liquid so formed would have in considerable degree the characteristic of a constant temperature mixture, such as the mixture of ice and water used to obtain a constant reference temperature. So long as any crystalline material containing quartz and the feldspars remained the temperature could not rise. Heat might flow freely into the mass from hotter surroundings but it would be consumed entirely in producing more granitic liquid, without significant rise of temperature. There would therefore be a long period during which only granitic magma existed and all the earliest intrusive material would be granitic. In many batholiths it would probably be the only intrusive material. In others, as time passed, all the more fusible materials would be melted and the temperature would rise as less fusible materials were taken into solution. The liquid would pass to granodioritic composition and perhaps on to dacitic. That it might attain dioritic composition can be entertained, but that it could ever reach the gabbroic seems quite beyond the bounds of credence.

Waters probably had the Sierra Nevada batholith in mind when he made his suggestion, and that batholith is almost continuous with the batholith of Southern California. Larsen's findings in the latter batholith show very poor agreement with the deduced picture of a batholith whose members had their origin in a process of refusion. The earliest intrusions were gabbroic; they were followed by tonalitic and later granodioritic intrusives. Finally there was some granite, but only a little.

There is no escape from agreement with Larsen that basaltic magma was the parental material and that the later intrusives were produced from it by fractional crystallization, with some modification by assimilation of foreign materials.

It is not necessary to go so far as to assert that the action pictured by Waters never occurs. It probably does. A general survey of batholiths might reveal how great importance it may have. Knopf's (1948) survey does seem to indicate that there may be batholiths entirely of granite. How often do composite batholiths have the order of intrusion expected of the selective fusion process? Larsen says that the order he finds is the normal order. In making this statement he probably has most prominently in mind the Mesozoic and Tertiary batholiths of the western cordillera. Some have suggested that refusion was a more prominent feature in batholith formation in early geologic times, especially in the pre-Cambrian. It may be that support for this concept could be found by considering them with the questions just asked in mind. Nevertheless one would require to exercise caution. Any indication that might be found of absence of all rock types except granite in ancient batholiths might easily be more apparent than real. Old terranes have ordinarily suffered much retrograde metamorphism. Of all the possible batholithic rocks the higher temperature varieties are the more susceptible to such changes and might easily be changed sufficiently to escape recognition as facies of the intrusive. Granite, on the other hand, already a comparatively low-grade rock, suffers less under these conditions and remains recognizable as such. It would require a very careful survey of all the facts to answer the question whether ancient batholiths are really significantly different from those of more recent times.

While we are examining these questions that grow out of the larger question of the parental nature of basaltic magma, it seems desirable to discuss one fact that has been regarded as fatal to the concept that granitic magma can be derived from basaltic by differentiation. This fact is the absence of granite or rhyolite in the Pacific basin, and, according to recent indications, in much of the Atlantic floor as well. The answer is inherent in Larsen's description, already quoted, of the development of the batholith of Southern California. As the liquid changed slowly by crystallization from basaltic to more salic composition it was from time to time pushed up from below by mountain-building forces. If we add to this concept the probability that gravitative settling of crystals is not sufficient to give separation of crystals and liquid, but that filter pressing comes into play during the periods of diastrophism, it is possible to suggest why there are no granites or rhyolites in ocean basins. The mechanics of true ocean floors are probably not such as to permit the formation of intrusive masses of batholithic dimensions in the first place. Only the dike feeders of lava flows may be developed. In addition the mechanics characteristic of continental diastrophism may never prevail in true ocean floors. Consequently the filter-pressing effect is not freely available for the production of granitic differentiates, but when island-arc mechanics develop rhyolites appear to be developed also.

The story told by the San Juan and the Southern California phenomena and the light it throws on other regions suggest that igneous action is a unity. The fashionable cleavage between plutonism and volcanism probably has no objective existence.

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COMPOSITION AND GENESIS OF PYROXENE AND GARNET RELATED TO ADIRONDACK ANORTHOSITE AND ANORTHOSITE-MARBLE CONTACT ZONES*

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Abstract

The genesis of some pyroxenes and garnets related to Adirondack anorthosite, anorthosite-marble contact zones and their metamorphic equivalent are discussed on the basis of 10 clinopyroxenes whose chemical analyses as related to optical properties have recently been presented by Hess, and upon one new orthopyroxene analysis and 7 new garnet analyses given here. It is found that the clinopyroxenes of the anorthositic rocks are high in alumina as compared with those of gabbro. In contact zones with marble, ferrosalite and grossularite-andradite skarn is developed on the marble side whereas an almandite-rich mafic gneiss is often found as a border facies on the anorthositic side. One sheet of salite gabbro is interpreted as the product of assimilation of ferrosalite skarn by gabbroic anorthosite magma.

Introduction and Acknowledgments

The author made a collection of Adirondack minerals for a systematic study of their chemical composition as related to the igneous rocks and their metamorphic equivalents in 1939, but other commitments during the war period necessitated deferment of completion of the project until the present time. The entire cost of the field expenses for the collection and the laboratory costs of the preparation of the mineral samples and of the chemical analyses was covered by grants from the Phillips Fund administered by the Department of Geology, Princeton University. Dr. H. L. James prepared the mineral concentrates for chemical analysis. All chemical analyses were made in the Laboratory for Rock Analysis at the University of Minnesota. Full details of the optical properties of the clinopyroxenes have been given by Hess (1949). A general account of the rocks of the Adirondack Mountains of Northern New York State has been previously published (Buddington, 1939).

Pyroxenes of Anorthositic Series

All the pyroxenes (Table 1) from the rocks of the anorthositic series, with the exception of the most mafic (5) are richer in alumina (4 to 7%) and ferric iron oxide (2 to 3%) than those (2.3 to 3.5 per cent alumina and 0.6 to 1.5 per cent ferric iron) in normal gabbro, norite and diabase sheets. The anorthositic pyroxenes are also in general slightly higher in lime and markedly lower in magnesia. The implications would seem to be that the pyroxenes of the anorthositic series crystallized in a different environment, either in a chemically different magma, under different

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* Princeton investigations of rock-forming minerals No. 5.

TABLE 1

CLINOPYROXENES FROM ADIRONDACK ANORTHOSITIC SERIES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	H_2O+	H_2O-	Other Con-stituents	Total
1	50.13	4.71	2.67	9.66	10.98	19.54	.81	.14	.58	.23	.91	.05		100.41
2	49.96	4.30	1.96	8.87	12.00	20.43	.52	.00	.67	.22	.64	.10		99.67
3	45.80	7.06	3.14	10.29	8.42	21.58	.68	.01	1.42	.20	.30	.10		99.00
4	46.85	5.03	2.91	13.63	7.03	21.90	.46	.00	1.47	.20	.24	.05		99.77
A	46.28	7.38	2.21	14.80	8.91	18.78	n.d.	n.d.	.59	n.d.	1.115			100.065
5	51.01	2.69	1.86	10.07	12.41	20.59	.47	.02	.40	.48	.19	.06		100.25
AVERAGE CLINOPYROXENE OF GABBRO AND NORITE FOR COMPARISON														
B	51.21	2.72	1.18	10.41	14.34	18.70	.29	.02	.73	.27	.16	.06	.09	100.18
C	50.39	3.95	1.73	9.20	15.15	17.82	.18	.09	.51	.19	.82		.07	100.10
ORTHOPYROXENE														
6	51.21	2.14	1.02	25.89	17.15	1.59	.10	.06	.28	.51	.16	.08		100.19
D	50.33	3.36	1.03	19.40	21.40	2.77			. 07	.71	1.14			100.21

physical conditions, or both. The nature of the differences would appear to be what might be thought to occur if the pyroxenes of the anorthositic series crystallized from a magma richer in the feldspathic constituents and poorer in magnesia than the normal gabbroic and noritic magmas. It was previously (Buddington, 1939, pp. 216–218, 235) concluded from field evidence that the anorthositic series was derived from a magma about equivalent to 85 per cent labradorite, 10 per cent pyroxene and five per cent of accessory minerals with a substantial quantity of volatiles. The hypothesis is therefore advanced here that the clinopyroxenes of the anorthositic series have a composition different from the clinopyroxenes of normal gabbroic and noritic rocks because they crystallized from a magma much richer in calcium and alumina and poorer in magnesia than the normal gabbroic and noritic magma.

Explanation of Table 1

(1) Augite from coarse Marcy type of anorthosite, road cut SW end of outlet of Lake Clear, St. Regis quadrangle. Analysts, R. B. Ellestad and Lee C. Peck. Corresponds to analysis 22, p. 655, Hess *Am. Mineral*, **1949.** From rock whose chemical analysis is given in *G. S. A. Mem.* **7,** No. 7, opp. p. 24.

(2) Augite from anorthositic gabbro, Keene valley road, 1 mile WNW of intersection with main N-S highway, Elizabethtown quadrangle. Analyst, R. B. Ellestad. Corresponds to analysis 21, p. 655, Hess *Am. Mineral.*, **1949.** From rock whose analysis is given in *G. S. A. Mem.* **7,** No. 25, p. 36.

(3) Salite from gabbroic anorthosite gneiss, quarry 0.3 mile east of Stickney Bridge, Ausable quadrangle. Analyst, R. B. Ellestad. Corresponds to chemical analysis 24, p. 656, Hess *Am. Mineral.* **1949.** From rock whose analysis is given in *G. S. A. Mem.* **7,** No. 10, opp. p. 24.

(4) Ferrosalite from gabbro pegmatite in Marcy type of anorthosite, quarry 3 miles east of Wilmington on south shoulder of Hamlin Mt., Lake Placid Quadrangle. Analyst, R. B. Ellestad. Corresponds to chemical analysis 26, p. 657, Hess *Am. Mineral.*, **1949.** From rock whose analysis is given in *G. S. A. Mem.* **7,** No. 26, p. 36.

(A) Ferrosalite from coarse pegmatitic Marcy type anorthosite. A. B. Leeds, **1878.**

(5) Augite from mafic gabbro dike. NE of schoolhouse, 2 miles SE of Gates Corners, Antwerp quadrangle. Analyst, Lee C. Peck. From rock whose analysis is given in *G. S. A. Mem.* **7,** No. 47, p. 52.

(B) Average of 10 augites from the gabbroic, noritic and diabase sheets of Stillwater, Mont.; Bushveld, S. Africa; Triassic diabase of eastern U. S.; Duluth, Minn.; and Pigeon Point, Minn. Analyses 9–12 and 28–33 inclusive, Hess, *Am. Mineral.*, **1949.**

(C) Average of 30 analyses of diallage pyroxene from the gabbro-norite family after P. Tschirwinsky, quoted in Schlossmacher K. Datensammlung gesteinbildender Mineralien, 1931. *Fortschr. Min.* Bd. **19,** T. 2, p. 253, **1935.**

(6) Hypersthene from norite, a sharply defined layer in border facies of anorthosite, 1.7 miles W of Gabriels, Saranac Lake quadrangle. Analyst, Lee C. Peck. From rock whose analysis is given in *G. S. A. Mem.* **7,** No. 24, p. 36. Sp. G. of hypersthene. =3.543. Composition equivalent to diopside 5.8, enstatite 41.4 ferrosilite 46.5 ($En_{47}Fs_{53}$).

(D) Hypersthene from coarse pegmatite, Marcy type. A. B. Leeds (1878). Composition equivalent to diopside 9.2, enstatite 49.6, ferrosilite 33.3 ($En_{60}Fs_{40}$).

Very coarse pegmatite—like development of clinopyroxene is found locally in a facies of anorthosite with a little associated sulfide and apatite. Balk (1944) has referred such developments to the effect of volatiles. The coarse clinopyroxene he described has indices of refraction similar to that of the pyroxene (1) of the coarse Lake Clear anorthosite.

The clinopyroxene (ferrosalite, Table 1, No. 4) of a gabbro pegmatite vein with sharp crustification growth and sharp walls in anorthosite, however, is substantially richer in ferrous iron relative to magnesia than the clinopyroxene of other members of the anorthositic series. This is in the direction of change that might be expected of a late residual magmatic fraction. However, the plagioclase (An_{40-51}) in the normal anorthosite and in the associated gabbro pegmatite (An_{48}) veins do not show the difference in composition which might theoretically be expected in such a case. The pegmatitic local segregation facies however is reasonably explicable as crystallization from local volatile—rich facies of the magma.

The augite (5) is from a rock, consisting of about two thirds mafic constituents of which one half is augite, which occurs as a dike in anorthosite. The composition of the plagioclase is similar to that of the anorthosite and the composition of the augite is also similar to the composition of the pyroxene (1) of the coarse anorthosite but slightly richer in ferrous iron and lower in alumina and ferric iron and therewith similar to the augite of normal gabbro. It differs from augite of normal gabbro, however in a high lime content.

Our present experimental data and theoretical principles are either not adequate or have not yet been used in the right way to explain the correlation of some mineralogical and rock relationships of the anorthosite series.

The orthopyroxene (6) comes from a sharply defined layer of norite within anorthosite, and is a ferrosilite—rich member of the hypersthene group. This is appropriate for a late stage facies of the anorthositic series such as it seems to be though the plagioclase is similar to that of normal anorthosite. A coarse hypersthene from the anorthosite analyzed by Leeds (*D*) is more enstatic as is a course hypersthene ($En_{79}Fs_{21}$) described by Balk (**1944**, p. 300).

PYROXENE AND GARNET IN METAMORPHIC CONTACT ZONES BETWEEN ANORTHOSITE AND MARBLE

There are numerous contact zones between anorthosite and included layers of marble in the Willsboro and adjacent part of the Ausable quadrangles. Usually there is a thin layered mafic gneiss with veined structure adjacent to the contact zone on the side of the anorthositic rocks and a skarn layer on the side next to the marble. The marble usually carries disseminated silicates. The pyroxene disseminated in the marble may vary from diopside to ferrosalite and the garnet disseminated in marble from grossularite to an andraditic grossularite. Scapolite and wollastonite may also be present locally. Quartz is often an associate of the garnet. Sphene may be present as an accessory mineral.

A contact metamorphic zone between gabbroic anorthosite (Whiteface type) gneiss and marble near Willsboro has been previously described (Buddington, 1939, p. 41–42; and Broughton and Burnham, 1944).

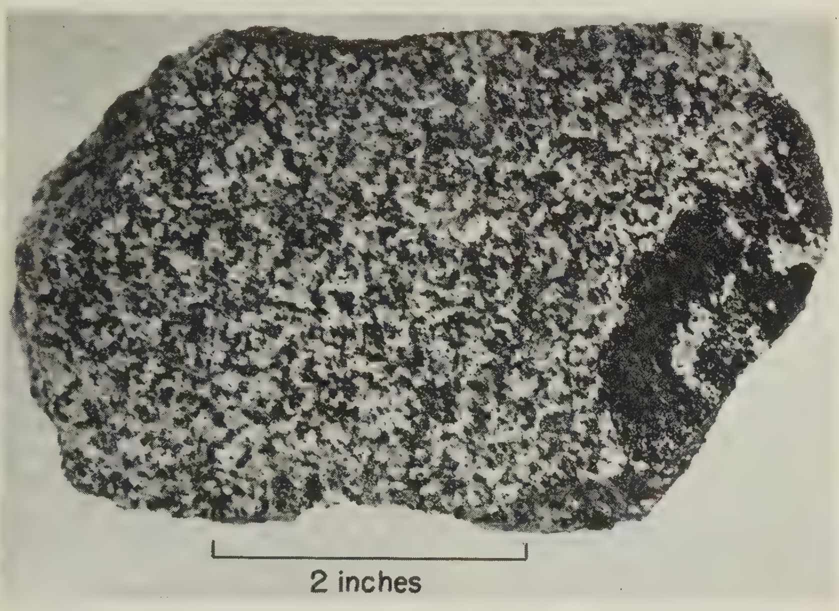

FIG. 1. Salite (near ferrosalite) gabbro, formed by assimilation of ferrosalite skarn (relic at right) in magma of gabbroic anorthosite composition. 3 miles SSW of Upper Jay, Lake Placid quadrangle.

Here the pyroxene from a skarn zone at contact with the anorthosite gneiss is a ferrosalite with the composition given in analysis 8. The high percentage of alumina is noteworthy.

At another locality, for about 3 miles SSW of Upper Jay on the Lake Placid quadrangle there is a synclinal sill of gabbro intruded into Grenville metasediments and containing random knots, shreds, and layers of ferrosalite skarn (Fig. 1). The bulk of the sill is more or less homogeneous but the rock has the appearance and texture of a member of the anorthositic series and its mafic character is attributed to disintegration of ferrosalite skarn with accompanying modification of the pyroxene (Buddington, 1939, p. 39). The pyroxene of the gabbro (contaminated anorthosite) is a salite (No. 9) near ferrosalite in composition. The skarn pyroxene and the pyroxene of the contaminated anorthositic sill are similar to some of the pyroxenes of the normal anorthosite, particularly that (4) of the pegmatitic facies.

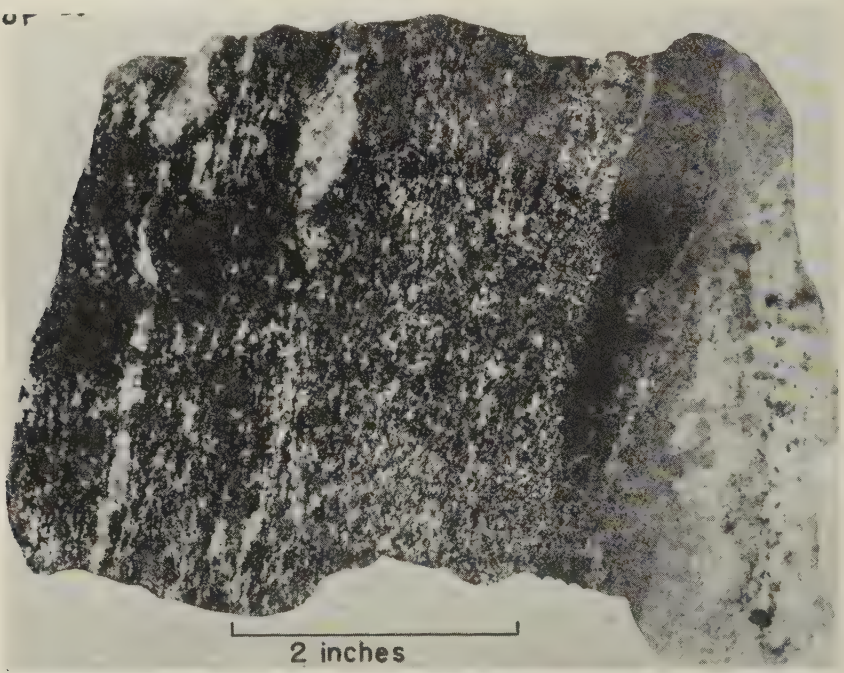

FIG. 2. Contact zone between gabbroic anorthosite gneiss and marble. Marble (light-colored) at right with disseminated andraditic-grossularite and ferrosalite. Mottled rock forming left 2/3rds of specimen is almandite-plagioclase-salite gneiss (mixed rock, mafic gneiss, formed from incorporation of ferrosalite skarn in gabbroic anorthosite). Narrow uniformly dark layer and thin light-gray layer between marble and gneiss is ferrosalite and andraditic grossularite skarn respectively. The two thin light-colored veinlets at left are younger granite pegmatite. Pokamoonshine quarry, Ausable quadrangle.

A contact zone between a mafic gneiss (a contaminated facies of the anorthosite) and an included layer of marble is well exposed on the floor of the old road material quarry near Pokamoonshine State Park, Ausable quadrangle. The marble layer is about 6 inches thick and carries disseminated ferrosalite (No. 7) and andraditic grossularite (No. 10). Between the marble and the mafic gneiss there is a narrow selvage consisting of andraditic grossularite skarn (No. 11) on the marble side and ferrosalite skarn next to the mafic gneiss. The mafic gneiss, about 2 feet thick, consists of plagioclase, salite (near ferrosalite) and almandite with a few thin granite pegmatite veinlets. (Fig. 2 is a photograph of a hand specimen from the contact zone). A short distance away there is a narrow layer of almandite-rich gneiss with subordinate plagioclase and pyroxene within the anorthosite which has the appearance as though pyroxene skarn were partly replaced by porphyroblastic plagioclase with subsequent development of almandite (Fig. 3). The garnet is an almandite (No. 12) and the plagioclase is oligoclase—andesine ($Ab_{70}An_{30}$). Vein-like masses of almandite are also found locally to cross the foliation of the meta-anorthosite as replacement veins, selvages of almandite develop locally on the border of mafic-rich layers, and it seems probable that at least part of the development of almandite is subsequent to the development of a mixed rock of ferrosalite skarn and plagioclase with plagioclase porphyroblasts.

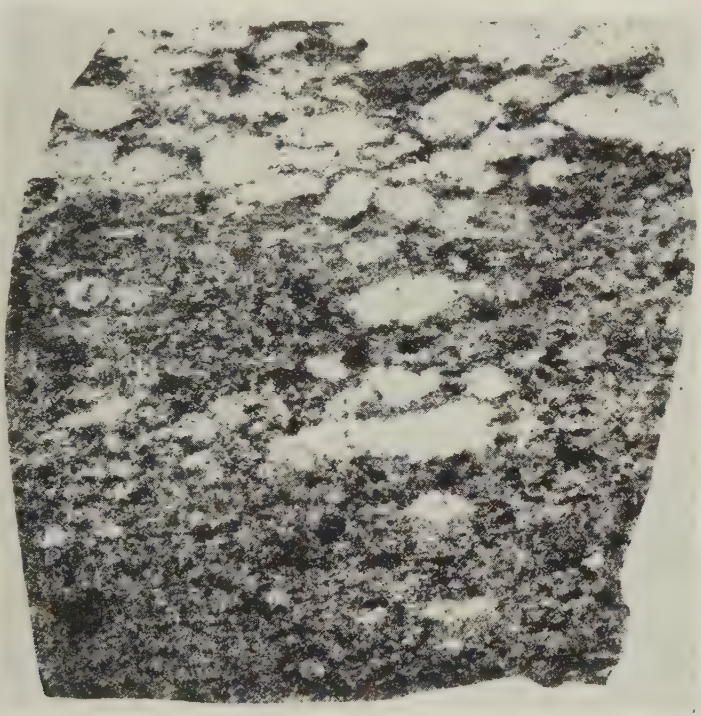

FIG. 3. Almandite-rich gneiss with subordinate pyroxene and plagioclase and augen of andesine. Andesine augen (white) are deformed porphyroblasts formed as replacement of ferrosalite skarn in contact zone between gabbroic anorthosite and marble. Almandite is later development, in part by reaction but in part with introduction of iron into mixed rock. Pokamoonshine quarry. Natural size.

TABLE 2

PYROXENES FROM CONTACT ZONE OF MARBLE AND ANORTHOSITE

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	H_2O+	H_2O-	S	Total
7	50.71	1.06	.53	18.57	5.70	22.86	.16	.02	.07	.18	.08	.04		99.98
8	45.30	7.74	3.36	12.27	7.04	21.74	.88	.16	1.30	.15	.39	.06		100.39
9	48.81	4.37	2.12	12.70	8.79	21.70	.53	.00	.70	.23	.10	.03		100.08
GARNETS FROM CONTACT ZONES BETWEEN ANORTHOSITE AND MARBLE														
10	38.57	18.88	5.22	2.27	.24	34.40	.00	.00	.34	.16	.04	.04	0.13	100.22*
11	38.57	17.54	6.76	3.00	.18	33.66	.00	.00	.42	.16	.05	.00		100.34
12	37.23	20.84	1.81	28.87	2.74	7.76	.00	.00	.17	.62	.08	.00		100.12
13	36.86	20.46	1.63	30.90	1.56	7.26	.09	.10	.16	1.13	.06	.04		100.25
14	37.30	21.04	1.01	29.29	2.77	7.37	.08	.06	.14	.93	.07	.03		100.09
15	37.23	20.91	1.40	28.37	2.13	9.01	.04	.02	.13	.92	.12	.04		100.32
16	37.48	21.01	1.12	27.25	3.48	8.45	.07	.03	.66	.82	.11	.05		100.53

* Total = 100.29
Less O for S = .07

Corrected total = 100.22

Another contact zone between gabbroic anorthosite gneiss and marble is exposed along the railroad about 0.5 mile south of the tunnel east of Rattlesnake Mt., Willsboro quadrangle. Here too there is a mafic veined-gneiss facies, about 2 feet thick on the anorthositic side of the zone and a thin skarn layer on the marble side. The garnet of the mixed rock is an almandite (14).

A contact zone between gabbroic anorthosite gneiss and an included

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Explanation of Table 2

(7) Ferrosalite from dissemination in marble layer enclosed in Whiteface type anorthosite gneiss, Pokamoonshine quarry, Ausable quadrangle. Analyst, Lee C. Peck. Analysis No. 16, p. 652, Hess *Am. Mineral.*, **1949.**

(8) Ferrosalite from feldspathic pyroxene skarn zone at contact with gabbroic anorthosite gneiss, 1.9 miles SW of Willsboro bridge, Willsboro quadrangle. Analysts, R. B. Ellestad and Lee C. Peck. Analysis No. 25, p. 657, Hess *Am. Mineral.*, **1949.** From rock whose analysis is given in *G. S. A. Mem. No.* **7,** No. 38, p. 44.

(9) Salite from gabbro, a facies of anorthosite mafic from incorporation of pyroxene skarn, quarry 2.5 miles SSW of Upper Jay, Lake Placid quadrangle. Analyst, R. B. Ellestad. Analysis No. 23, p. 656, Hess *Am. Mineral.*, **1949.** From rock whose analysis is given in *G. S. A. Mem. No.* **7,** No. 30, p. 36.

(10) Grossularite (andraditic) from dissemination in central part of a one-foot-thick marble layer enclosed in Whiteface type anorthosite gneiss, Pokamoonshine quarry, Ausable quadrangle. Analyst, Lee C. Peck. Sp.G.=3.652.

(11) Grossularite (andraditic) from massive garnet and pyroxene skarn layer at contact between mafic anorthosite gneiss and marble, same locality as 10. Analyst, Lee C. Peck. Sp.G.=3.670

(12) Almandite from gneissic feldspathic pyroxenic almandite layer several inches thick enclosed within Whiteface type anorthosite gneiss. Same locality as 10 and 11. Analyst, Lee C. Peck. Sp.G.=4.035. Plagioclase is oligoclase—andesine ($Ab_{72}An_{28}$).

(13) Almandite from ultramafic pyroxene layer (? skarn) several feet thick enclosed within gabbroic anorthosite gneiss, road cut about 2 miles NW of Willsboro, Willsboro quadrangle. Analyst, Lee C. Peck.

(14) Almandite from a metamorphosed migmatite consisting of alternate laminae of pyroxene-garnet skarn and anorthosite, forming a 2-foot contact zone between anorthosite gneiss and marble, ½ mile south of R. R. tunnel east of Rattlesnake Mtn., Willsboro quadrangle. Migmatite has percentages of minerals as follows; plagioclase 28, almandite 26, potash feldspar 6, quartz 6, salite 12, hornblende 14, magnetite 6, and apatite 2. Analyst, Lee C. Peck.

(15) Almandite from metamorphosed migmatite of gabbroic anorthosite gneiss and schlieren and shreds of pyroxene-almandite skarn in contact zone between marble and anorthosite. The marble is a layer included within anorthosite. Locality is 2½ miles NE of Cross, Ausable quadrangle, road cut just south of intersection of side road and main highway. Migmatite has about the following percentages of minerals, andesine 60, microcline 5, quartz 0.5, almandite 18, pyroxene 8, hornblende 3, magnetite 4, and apatite 1.5. Some sphene is present. Chemical analysis of rock given in *G. S. A. Mem.* **7,** Table 10, No. 40. opp. p. 44.

(16) Almandite from sparsely-disseminated porphyroblastic garnets in gabbroic meta-anorthosite gneiss, road cut just east of outlet of Long Pond, Willsboro quadrangle. Analyst, Lee C. Peck.

layer of marble has previously been described (Buddington 1939, No. 40, Table 10) from near Cross, Ausable, quadrangle. The mafic border facies of the anorthositic rock has the composition of a diorite and the structure of an arteritic migmatite comprised of anorthositic veinings or impregnations and skarn. Garnet forms about 18 per cent of the rock and is an almandite (No. 15).

The garnet from an ultramafic pyroxene layer within gabbroic anorthosite gneiss about 2 miles NW of Willsboro and a disseminated sparsely porphyroblastic garnet from gabbroic anorthosite in the same belt have been analyzed (Analyses 13 and 16). Both are almandite.

The clinopyroxene of the almandite-rich mixed rock of anorthosite veins and impregnations and pyroxene skarn in general is uniformly a salite with optical properties very similar to that of the salite (9) from the Jay gabbro (anorthosite contaminated by skarn) and represents a modification of the original skarn ferrosalites (7 and 8) whereby there is an increase in the ratio of magnesia to ferrous iron.

It was originally expected that the garnet concentrated on the anorthositic side of the contact with skarn would be found to have a direct genetic connection with the garnet of the skarn since both were related to the

Table 3. Garnet Analyses Recalculated to Equivalent Molecules

	Ratio Mol % MgO/FeO in Rock	Almandite	Pyrope	Grossularite	Andradite	Spessartite	
				In Meta-limestone			
10		4.5	0.8	77.4	16.8	0.5	Pokamoonshine, Ausable quadrangle.
11		6.6	0.6	67.7	24.6	0.5	Pokamoonshine, Ausable quadrangle.
				In Meta-anorthosite Veined Pyroxene Skarn Zone			
12		63.4	8.6	21.6	4.3	2.1	Almandite—rich layer in anorthosite-skarn migmatite, Pokamoonshine.
13		70.0	6.4	16.3	4.9	2.4	Corona garnet in pyroxene skarn, Willsboro.
14		66.3	11.2	17.4	2.9	2.1	Anorthosite-pyroxene skarn migmatite, Rattlesnake Mtn.
15	25/75	63.4	8.6	21.6	4.3	2.1	Anorthosite-skarn migmatite, Cross.
				In Meta-anorthosite Gneiss			
15	22/78*	60.3	13.9	20.7	3.3	1.8	Porphyroblastic garnet, Long Pond, Willsboro.

* Average of composite sample of similar rock.

contact zone. This has not proven to be the case. The garnet in the normal anorthositic gneiss and its local mafic contact facies has uniformly proven to be almandite, whereas the garnet in the skarn or disseminated in the marble equally uniformly belongs to the grossularite—andradite series. The concentration of almandite in such mixed rock is a local phenomenon and does not always occur in all mixtures of anorthosite-veined pyroxene skarn, though it is common.

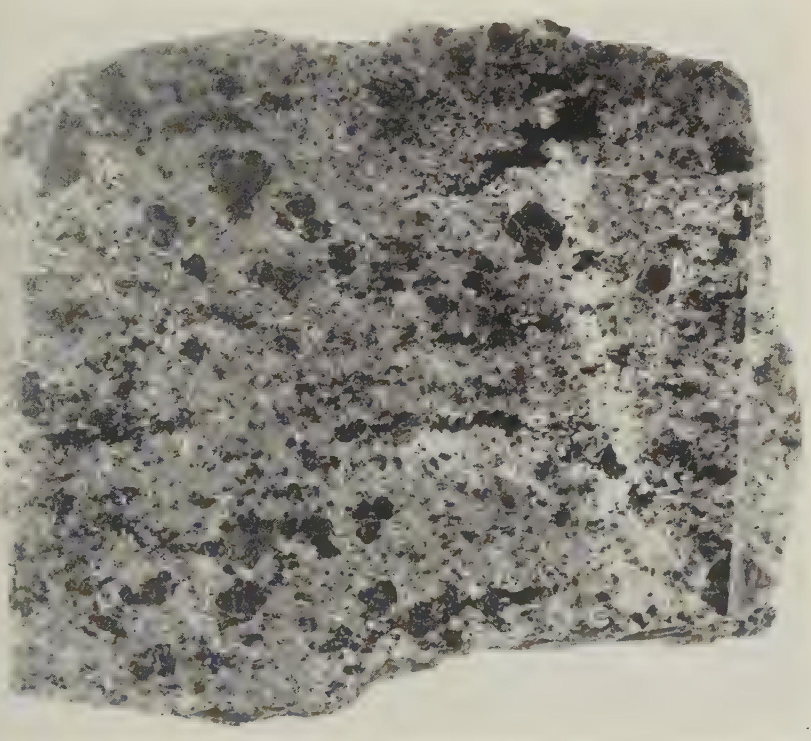

FIG. 4. Porphyroblastic almandite (large dark grains) in gabbroic anorthosite gneiss. Largely a regional metamorphic reaction product. East of outlet of Long Pond, Willsboro quadrangle.

It is believed that the following sequence of events will best serve to explain the minerals of the contact zones. (1) In the first stage of emplacement of a gabbroic anorthosite magma, disseminated diopside and ferrosalite developed consequent upon silication of the adjoining marble, and ferrosalite skarn formed adjacent to the magma itself; at least part of the grossularite-andradite may also have developed at this time; (2) the ferrosalite skarn was then intruded and impregnated by anorthositic materials, somewhat modified in composition, and accompanied by porphyroblastic development of andesine in the skarn layer; (3) iron rich solutions especially affected the contact zones developing almandite in the mixed anorthosite—skarn rock and perhaps adding to the andradite-grossularite on the skarn and marble side. Periods 2 and 3 may well have overlapped; (4) the border zone of the great anorthosite massif as a whole together with its local mafic contaminated border facies was strongly deformed and underwent plastic flowage with the development of some corona almandite and the recrystallization and regrowth of prophyroblastic almandite. This was followed by injection of granite pegmatite veinlets and veins which in turn locally carry almandite.

Coronas of almandite at the contact between pyroxene and plagioclase are found in much of the feldspathic skarn and in migmatite and permeation mixed rock of skarn and anorthosite, so that the possibility may be considered that all the almandite has had an origin through reaction during regional metamorphism. Doubtless part of the garnet did have this origin. This would require substantial leaching of lime and silica together with some magnesia. The plagioclase of the very strongly garnetiferous rock is oligoclase-andesine ($Ab_{70}An_{30}$) in contrast to the normal andesine-labradorite of the anorthosite, and consistent with this hypothesis. However, there is so much iron in the almanditic mixed rock that it would seem highly probable that some had been introduced. Local prehnitization of the plagioclase in the contact zones may be related to the movement of lime by hydrothermal solutions which effected the development of the almandite.

It seems equally probable however that the disseminated porphyroblastic garnet of the normal meta-anorthositic facies (Fig. 4) has developed from reaction between ferrosalite and plagioclase with but little modification in bulk chemical composition.

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CAMPTONITE DIKES NEAR BOULDER DAM,[1] ARIZONA[2]

IAN CAMPBELL* AND EDWARD T. SCHENK**

ABSTRACT

Near Boulder Dam occur camptonite dikes of possible Quaternary age characterized by amphibole (kaersutite) phenocrysts of exceptional size and unusual distribution—a distribution which suggests that these phenocrysts have formed essentially in place. Rough calculations suggest that these crystals grew to four-inch size in less than 25 days. Serpentinization of olivine, formation and size of amygdules, and growth of amphibole phenocrysts are related to distribution of volatiles in the dike magma. Chemical analyses of the amphibole and of the chilled margin of a dike are presented and discussed.

INTRODUCTION

Camptonite dikes occur within a limited area bordering U. S. Highway 93, approximately eight miles south of Boulder Dam, on the Arizona side. Several unusual features combine to make this occurrence unique:

1. Wall rocks: these are fan gravels, not greatly consolidated; not perceptibly metamorphosed.
2. Texture: the dikes exhibit a striking porphyritic texture, dominated by amphibole phenocrysts of exceptional size (up to four inches) and uncommon character.
3. Composition: the rock is a camptonite and the amphibole most nearly resembles kaersutite. Rocks with such alkaline affinities have been unknown in this region.
4. Age: the dikes are probably late Pleistocene or Sub-Recent.

The locality, lying as it does along a major highway, has been seen by many geologists and the exposures are sufficiently striking to attract the attention of even the casual tourist. Specimens have found their way into many collections, both public and private. Hence a report on field and laboratory studies, carried on at intervals over a period of several years, may be of interest.

The area (see Fig. 1), lies in the northwestern part of the old (1884) Camp Mojave (Arizona-California-Nevada) quadrangle of the U. S. Topographic Atlas. There is no published geological information giving

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[1] In 1947 the name "Boulder Dam" was changed by act of the 80th Congress to "Hoover Dam." Inasmuch as our studies were initiated and our maps completed during the earlier regime, we have retained the name by which this structure was then known.

[2] Contribution of the Division of the Geological Sciences, California Institute of Technology, No. 539. Published with the permission of the Director, National Park Service.

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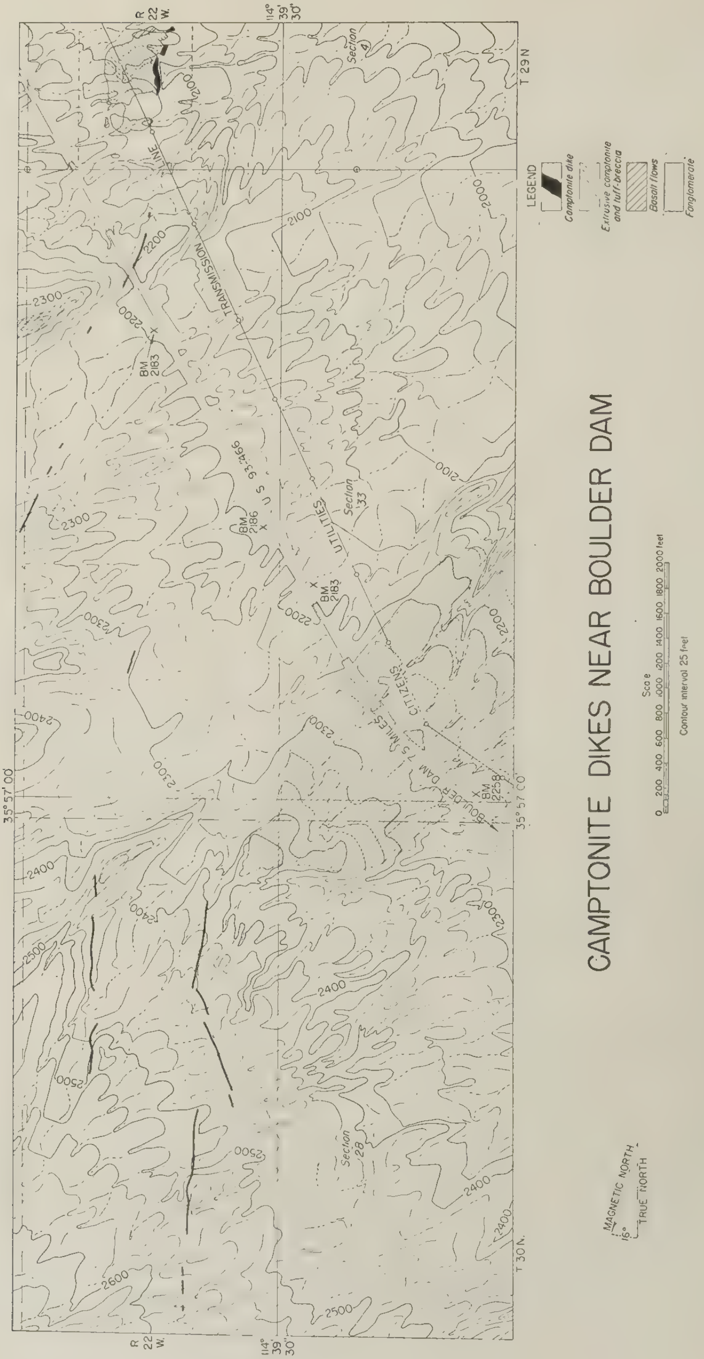


CAMPTONITE DIKES NEAR BOULDER DAM

FIG 1.

any detail on this region. Lee (1908), does not discuss the district, but does represent it on the map accompanying his report as an area of Tertiary basalt, flanking a preCambrian basement. Darton (1924) on the state map of Arizona shows the area as undifferentiated Tertiary (including possible Quaternary) and older volcanic rocks. Longwell (1936), in his excellent study of the geology of the floor of the Boulder reservoir, has summarized the literature dealing with the general region; but neither his report, nor that of Ransome (1923, 1931) extends far enough south to cover the tract under discussion.

The Geologic Setting

Fanglomerate

The oldest and most wide-spread formation in the area under consideration, is a deposit at least 100 feet thick of fanglomerate. This formation slopes southwesterly from the Black Mountains and extends some ten miles, practically to the Colorado River below Boulder Dam. The fanglomerate (see Fig. 2) at first sight gives the impression of having been formed relatively recently, but the deep insequent canyons incising the fan surface and the basalt-capped residuals suggest that the principal depo-

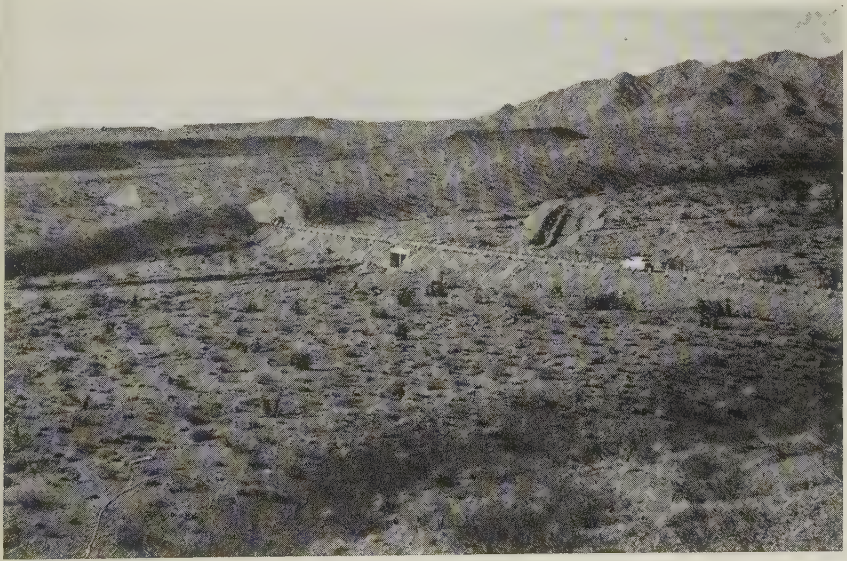


Fig. 2. Looking northeast across U. S. Highway 93 about eight miles south of Boulder Dam, Arizona. The Black Mountains are in right background; flows of olivine basalt form the dark cappings in left middle background and to right of center; fanglomerate sections are exposed by the highway excavations.

sition of the fanglomerate may belong to a somewhat earlier cycle than the present.

The bulk of the fanglomerate has been derived from "Basement Complex," and fragments of a wide variety of rock types are present. These include granite, gneiss, schist, amphibolite, pegmatite, and vein quartz. Locally there has been a slight amount of cementation by a calcareous "caliche," but in general the fanglomerate is very little indurated.

Olivine Basalt

Within the mapped area and also in the surrounding region, flows of olivine basalt are found intercalated with and overlying (see Fig. 2) the fanglomerate. These flows are not thick, the maximum noted being twenty feet. Specimens from different flows are lithologically almost identical. The rock is dark grey to black, very fine-grained, and dense for the most part, although small vesicles (filled and unfilled) occur, particularly near the top and bottom portions of flows. Thin-sections reveal that the rock is a typical basalt consisting of abundant narrow laths of median labradorite, averaging between 0.1 and 0.2 mm. in length and distributed in a microaphanitic groundmass. This groundmass contains numerous tiny euhedral grains of magnetite, less abundant pyroxene, occasional olivine, and attendant alteration products. Microphenocrysts of olivine, up to 2.0 mm. in diameter, occur sparsely in the rock, and millimeter-sized glomeroporphyritic aggregates of pigeonitic pyroxene are sometimes found. Labradorite is commonly quite fresh; the larger pyroxenes show a small amount of resorption rounding; olivine shows incipient to moderate serpentinization. Flow texture is frequently well exhibited by a subparallel arrangement of the feldspar laths.

Camptonite

The most interesting rock within the area and the one with which this paper is chiefly concerned is an intrusive camptonite found as dikes in the fanglomerate. By far the best exposure of this camptonite is in a road cut along Highway 93, 7.5 miles south of Boulder Dam. The vertical exposure here (see Fig. 3) is about twelve feet, and within this range the dike maintains a width of four feet between parallel, very straight, and nearly vertical walls. Southward from the cut the dike can be traced by small surface outcrops and residual talus for a distance of almost 2000 feet, and to the northeast for a slightly greater distance. Some three thousand feet north from the road cut, exposures occur of similar dikes, which may or may not represent bifurcations of the first. It may be significant, or it may be no more than a curious coincidence, that a float specimen, lithologically identical with these dikes, was found about thirteen miles distant near Callville Wash north of Lake Mead and in line with the general northerly trend of these dikes.

That the dikes intrude the fanglomerate seems clear; yet to some geologists the occurrence of well-defined dikes in fan gravels has seemed so astonishing that they have suggested that the camptonite was originally intruded into some older formation which has since been removed by differential erosion, and that the fanglomerate was subsequently laid down around the dikes. In other words, this hypothesis would have the present contact of fanglomerate and dikes a depositional one rather than intrusive. Certainly no one would question the intrusive character of the dikes themselves: their pattern on the map; their vertical walls; their chilled margins—all point to typical intrusive structures. The problem is therefore whether or not the present fanglomerate was the country rock at the time of intrusion. If the dikes had been intruded into an earlier formation, then it would seem reasonable that some small residual of this should have been discovered; but in our examination of the several thousand feet of exposures of these dikes nothing like this has come to light, nor is any such hypothetical formation known from the section, although our knowledge of this is admittedly scant. If the dikes had been left standing as buttresses to be filled around by the fanglomerate, some depositional "lee and stoss" features should be apparent in the

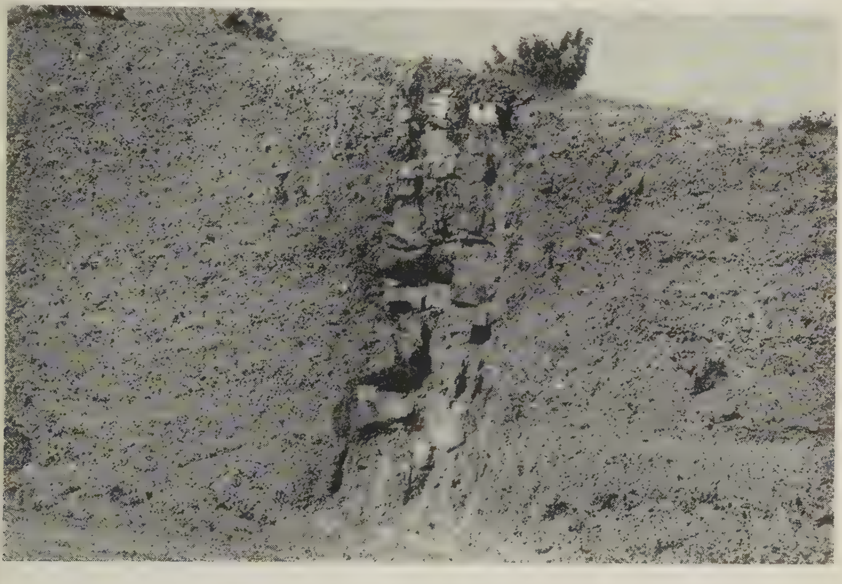


FIG. 3. Section of camptonite dike in fanglomerate, exposed in road cut on U. S. Highway 93, 7.5 miles south of Boulder Dam, Arizona.

sediments. Yet in the excellent cross section provided by the highway cut, the same sedimentation units are present in the fanglomerate on both sides of the dike. These points are, in a sense, negative evidence; unfortunately there is little positive evidence of intrusion of the dikes into the present fanglomerate. The dikes commonly show sharp and surprisingly smooth contacts against the fanglomerate. Although no distinct apophyses were discovered, in a few instances minor protrusions from walls of the dikes have bulged around cobbles in the fanglomerate in a manner highly unlikely to result from depositional contact of gravel against a residual dike. In certain favorable lights a faint zone of reddening is apparent at the highway cut extending for a distance of eight to ten feet into the fanglomerate equally on either side of the dike, and clearly related to the intrusion. Our conclusion is that the dikes are intrusive into and younger than the fanglomerate.

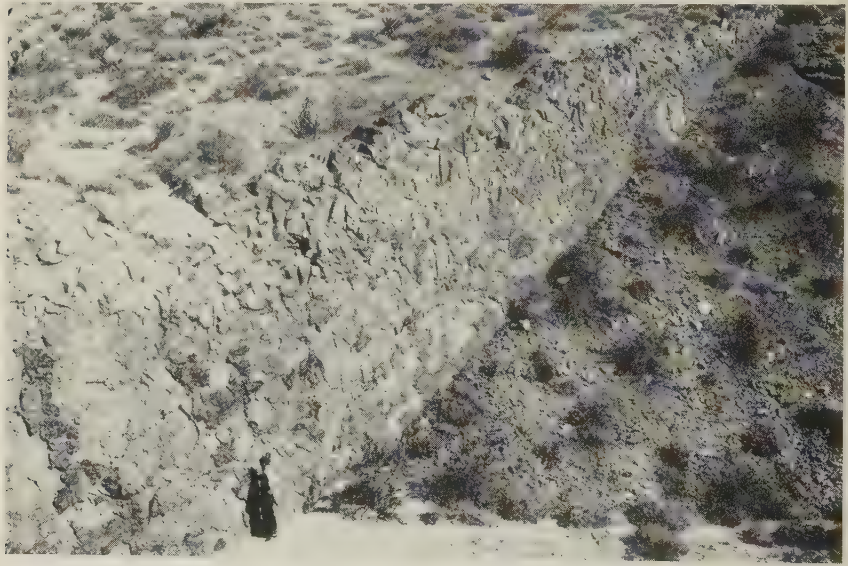


Fig. 4. Downstream face of "dam" formed where camptonite dike crosses a dry wash in fanglomerate, in northern part of map area, along central segment of the three northwesterly striking camptonite outcrops shown in Fig. 1.

The age relations between the camptonite dikes and the basalt flows are obscure. Nowhere have the two formations been discovered in actual contact. Meager evidence, in the form of basalt fragments included in a camptonite breccia (discussed beyond) suggests that the camptonite is younger than at least some of the basalt.

Throughout most of the exposures of these dikes, the same general features and relationships persist. Widths vary from two to six feet, but average close to four for most of the dikes; dips vary from 65° E to 75° W, but average close to vertical; strikes vary from N 20° W to N 25° E, but average nearly north. Commonly the dikes are somewhat more resistant than the fanglomerate and therefore appear as ridges. In a few places, where a dike crosses a wash, a notable "dam" has developed (see Fig. 4). It is possible that the offsets appearing at several points along the course of the dikes (Fig. 1) are due to faulting, but in the absence of marker beds in the fanglomerate, this cannot be demonstrated. The offsets might equally well represent sharp inflections in strike of a dike, coupled with a pinching in the conduit; that is to say, these "segments" may be connected and continuous in depth.

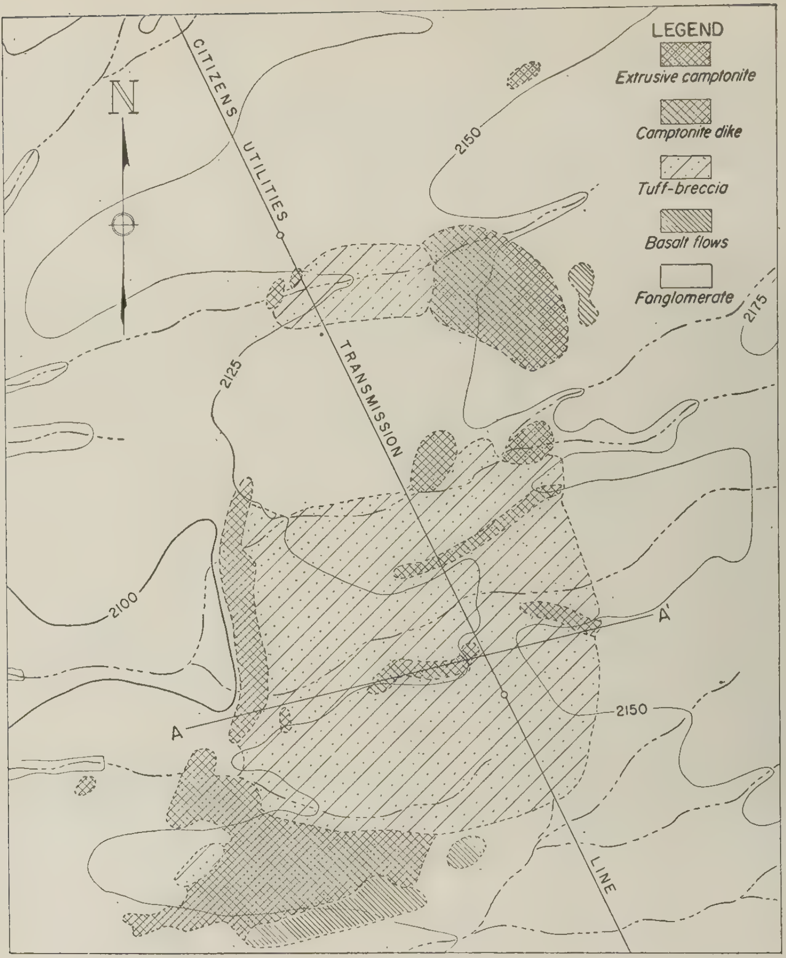
In the southern part of the map area, where outcrops of the camptonite terminate, complex relations exist which cannot be entirely resolved from the limited exposures. Because of the presence here of pyroclastic facies, not elsewhere recognized, and of sill-like structures of the camptonite, the suggestion is strong that at this point there may have been a focus of the volcanic activity associated with the intrusion of the dikes. From detailed mapping of the limited exposures (see Fig. 5), the most probable chronology of events appears to be:

First: deposition of the fanglomerate.

Second: flow of basaltic lava, source unknown.

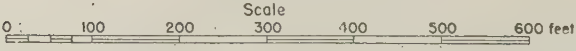
Third: development of a small vent (position is indicated on section $A-A'$, Fig. 5). Largely filling the vent is a pyroclastic formation, mapped as "tuff-breccia," which varies from a brown tuffaceous rock carrying small crystals of amphibole identical in its optical properties with phenocrysts of the dikes to a coarse breccia containing larger amphibole crystals, "bread-crust bombs" (one to two feet in diameter) of camptonite, and abundant material of accidental origin, such as pebbles from the fanglomerate, and some large (two foot) fragments of basalt. The texture of this formation coarsens rapidly from east to west, suggesting that the source lay to the west in the direction of the assumed vent. Since the base of this formation is not exposed, the thickness is unknown, but presumably not more than a few tens of feet are involved.

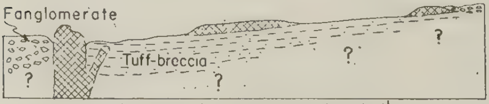
Fourth: intrusion of camptonite magma along the western margin of the supposed vent, where, having pushed through to the surface, it appears also as an extrusive facies overlying the tuff-breccia. This extrusive facies is characterized by abundant amygdules with horizontal or gently dipping elongation. The possibility has been recognized that the "extrusive camptonite" might actually be a sill with the roof now completely stripped. However, since it overlies sub-aerial pyroclastics and elsewhere also overlies fanglomerate, the possibility of its being a sill seems remote.



MAP OF

THE AREA* AT THE SOUTHERN END
OF THE CAMPTONITE DIKES





Generalized section along line A-A'

*Outlined by dashed-line rectangle within figure 1.

FIG. 5.

Petrology of the Camptonite Dikes

The only well-exposed cross section of a camptonite dike is in the highway cut (Fig. 3) described above. The following discussion therefore pertains principally to this exposure. Texture is the most striking feature of this dike. Distributed through a grey, aphanitic groundmass are abundant, large, glisteningly fresh, black crystals of amphibole some of which measure as much as four inches in greatest diameter. At first sight most laymen and some geologists have identified these amphibole phenocrysts as obsidian fragments (see Fig. 6). The resemblance is indeed close, for the amphibole cleavage is not often apparent, and resorption rounding on some crystal surfaces has simulated the conchoidal fracture of obsidian. In addition to the amphibole phenocrysts, rust-colored pseudomorphs of olivine one-half inch or less in diameter occur occasionally. Amygdules are common, the smaller ones completely and the larger ones partly, filled with calcite. Inclusions are absent, except for a few small aggregates of quartz which appear to be of accidental origin.

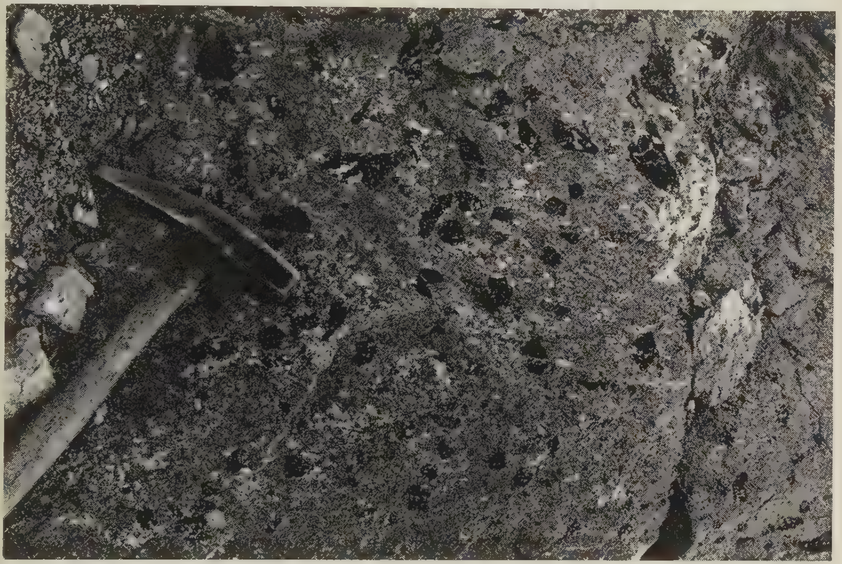


Fig. 6. Large black phenocrysts of amphibole in displaced block of camptonite. Note increase in size of phenocrysts from bottom margin of photograph toward the top. Bottom of photograph represents portion of block nearest to wall of dike. Small white areas are calcite and calcite-analcime amygdules and veinlets.

The distribution of phenocrysts and of amygdules within the dike is at once interesting and significant. A chilled selvage of aphanitic rock, in which amygdules and amphibole phenocrysts are practically absent, makes up the outer four to six inch margin along both walls of the dike. Inside of this selvage, amphibole phenocrysts and amygdules appear, increasing progressively both in size and in number toward the central portion of the dike. This change is distinctly gradational and rather symmetrical with respect to both walls of the dike. As a check on what would otherwise be only a strong visual impression, an attempt was made to get a quantitive evaluation of this textural change with respect to the amphibole phenocrysts. If a large plane-surfaced cross section had been available, a megascopic Rosiwal analysis might have been possible; but lacking this, the most practical method and one sufficiently rigid to exclude personal bias seemed to be to mark off convenient vertical zones in the dike and measure the maximum diameter of each hornblende phenocryst within an arbitrarily drawn cross section. Between 25 and 50 crystals were measured in each of seven zones, numbered in sequence from the selvage on the east or west wall of the dike, toward the center. Thus Zone $\mathrm{I_e}$ (see Table 1) is the first zone beyond the chilled border adjoining the east wall of the dike; Zone $\mathrm{I_w}$ is the corresponding zone adjoining the western chilled margin. The figures immediately below the zone number give the distance in inches from the wall of the dike.Measurements were made only to the nearest tenth of an inch, consequently the second decimal given in the figures for average diameter is perhaps more of arithmetical interest than of statistical significance. Krumbein (1935) has shown and Chayes (1950) has recently confirmed the fact that "the trace of any grain dimension along a section of the grain will usually be shorter than the true dimension." This is recognized in the presentation above, but relative dimensions are here of more significance and these are probably of the right order of magnitude. At the same time, the absolute size of the amphibole phenocrysts deserves mention. They

TABLE 1

Zone:	Chilled border	$\mathrm{I_w}$	$\mathrm{II_w}$	$\mathrm{III_w}$	$\mathrm{IV_{ew}}$	$\mathrm{III_e}$	$\mathrm{II_e}$	$\mathrm{I_e}$	Chilled border
Distance from wall of dike (inches):	0–6	6–10	10–14	14–20	20–26	20–14	14–10	10–6	6–0
Average maximum diameter of amphibole phenocrysts (inches):	absent	0.27	0.46	0.54	1.38	0.61	0.53	0.31	absent
Range in diameter:	—	0.1–0.5	0.2–1.0	0.3–1.1	0.5–2.5	0.2–2.1	0.2–1.7	0.2–1.6	—

Average maximum diameter for entire section: 0.59 inches.

range up to four inches in maximum diameter, an astonishing figure for this mineral in this kind of rock. Palache (1932) lists the largest known hornblende as 15×15×46 cm. This was a hastingsite from a pegmatite, a very different type of occurrence from what we are here considering. Feniak (1944) in his study of grain dimensions in igneous rocks gives the average cross-sectional area for hornblende as only 0.42 sq. mm., with the average long dimension as 0.8 mm. Although no exhaustive search of the literature has been made, it seems likely that these 4-inch crystals are among the largest amphibole phenocrysts known from camptonite and related igneous types.

The amygdules show the same general pattern of size distribution as do the amphibole phenocrysts. They show also a moderate degree of alignment with the longer axes paralleling the walls of the dike. No such flowage texture is exhibited in the case of the amphibole phenocrysts which show on the whole a nearly random orientation. Unlike amphiboles and amygdules, the rusty subhedrons of altered olivine are distributed evenly although sparsely throughout the dike, *including the chilled margins.*

A series of six thin-sections cut at intervals from a two foot wide specimen of the dike afforded detailed comparisons across a section from chilled margin to center. The microscopic data presented is based upon these together with various random sections cut from this and other dikes in the area.

The groundmass is in considerable part microaphanitic and shows little change from wall to center of dike. Microlites of calcic andesine, usually less than 0.1 mm. in length, are abundant, as are tiny euhedrons of magnetite and somewhat larger subhedrons of olivine largely altered to serpentine and "limonite." Apatite is a rare accessory. Distributed rather abundantly, even in the groundmass of the dike margins, are small vesicles completely filled with calcite, or less commonly with calcite and analcime, and still more rarely with analcime alone. Where these two minerals occur together, analcime appears to have been the earlier. Analcime is also somewhat more common in the larger amygdules. The vesicles are usually ovoids, ranging in size from 0.25 mm. to several centimeters in the case of the larger amygdules near the center of the dike. A few vesicles are highly irregular in shape and some pass into calcite or calcite-analcime veinlets. The elongation, like that of the feldspar microlites, parallels the walls of the dikes.

In addition to the large crystals of amphibole a number of other minerals are present as phenocrysts and microphenocrysts: olivine, diopside, labradorite, sanidine, and quartz. With the exception of olivine these are all rare; seldom are more than one or two grains present in a single thin-section. All except olivine are fresh. All, again excepting olivine, show considerable resorption rounding, but only quartz shows any reaction border, and this is of indeterminate character. Quartz in some instances consists of composite grains, and it is not improbable that these are xenocrysts rather than crystallizations from the dike magma.

In thin-section, just as in hand specimen and in the field occurrence, the amphibole phenocrysts are the most striking constituent. Under the microscope these crystals appear exceptionally fresh. Occasionally they may be veined by calcite especially along opened cleavages, but this is in the nature of an open-space filling rather than an alteration. Indeed, in the larger crystals it is often possible to split a layer of calcite from a cleavage plate of the amphibole leaving it entirely free of the carbonate. Indices of the amphibole are: $\alpha = 1.670$; $\beta = 1.692$; $\gamma = 1.701$. Extinction angle ($c \wedge Z$) is 8° to 10°; $2V$ (−) is 81°. Crystals are strongly pleochroic with X = light yellow brown, Y = reddish brown, Z = dark reddish brown, and $X < Y < Z$. Measurement of alpha and gamma indices and of extinction angles* on a series of amphibole phenocrysts spaced at intervals across the width of the dike show no differences beyond such as fall within errors of measurement in this group (Turner, 1942). Phenocrysts in thin section show uniform and sharp extinction throughout, save for a very few in which an outer rim (less than 0.1 mm.) has an extinction angle greater by 2° to 3° than that of the main body of the crystal. Therefore, it is concluded that no significant compositional changes are involved, either within individual phenocrysts or with respect to phenocrysts in different zones within the dike.

In color, in high refringence and birefringence and in strong pleochroism this amphibole resembles basaltic hornblende or lamprobolite (Rogers, 1940). Winchell (1945) in codifying the properties of the oxyhornblendes (another name for this type of amphibole), pointed out that the size of the extinction angle is a guide to the amount of oxidation (the smaller the angle, the greater the oxidation). Hence it would seem that this amphibole is one in which relatively little oxidation has taken place as compared to the majority of this group for which extinction angles are commonly less than 5°.

Indeed despite much excellent work on the subject, the nomenclature of amphiboles is still unsatisfactory. As pointed out by Larsen and Berman (1934, p. 220) nomenclature has depended more on optical peculiarities than on the more fundamental feature of chemical composition. An analysis of this amphibole is given in Table 3, column *C*. This corresponds approximately to a formula of $Ca_4(Na,K)_2(Mg,Mn,Fe^{++})_7$

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* To facilitate obtaining favorable orientations, use was made of "grain thin sections" (Von Huene, 1949) prepared from fragments chipped from phenocrysts in the dike.

$(Al,Fe^{+++})_6TiSi_{12}O_{46}(OH,F)_2$. The relatively high content of titanium is a significant feature, and because of this, coupled with its other characters, Berman (1940) suggested *kaersutite* as the best available designation for this amphibole. In terms of compositional relationships, it is of interest to point out that the analysis of this amphibole places it just on the border (toward CaO and adjoining the pyroxene field) of the amphibole field in Kennedy's (1935, Fig. 2) diagram. Hallimond (1943) has developed a graphical presentation of significant atomic ratios in amphiboles and has delimited thereon the fields of amphiboles from various rock types. Data for this amphibole, when plotted on Hallimond's chart, place it in a vacant triangle (close however to Hallimond's analysis No. 143) which lies in a zone where the fields of basalt, diorite, and nepheline syenite approach or overlap. This is consistent with other evidence, which has indicated the basaltic and alkaline characteristics of the camptonite in which this amphibole occurs.

CLASSIFICATION

It is impractical to obtain a volumetric mode for this rock because of the very fine grain and indeterminate character of part of the groundmass and because of the extreme variation of grain size and variable distribution of some constituents. A summary of the mineral components, listed in approximate order of decreasing abundance, is given in Table 2.

TABLE 2. MINERAL COMPOSITION OF BOULDER DAM CAMPTONITES

Plagioclase (calcic andesine): microlites in groundmass.
Olivine (largely altered to serpentine and "limonite"): phenocrysts and groundmass.
Amphibole: (kaersutite) phenocrysts.
Calcite: confined to amygdules and veinlets.
Magnetite: groundmass.
Diopside: phenocrysts.
Labradorite: phenocrysts.
Quartz: xenocrysts (?).
Analcime: confined to amygdules and veinlets.
Sanidine: phenocrysts.
Apatite.
Zeolites (small amounts of an as yet unidentified zeolite may be present along with analcime in some of the amygdules).

This assemblage clearly places the rock in Johannsen's (1937) family No. 3216H (Camptonite) and in practically all respects the rock corresponds to the usual definitions of camptonite.

A chemical analysis of this rock is given in Table 3 column *A*, and in column *B*, for comparison, is Daly's (1933) average of 15 camptonites.

The analysis was made upon material taken from chilled margins of the dike, since this, as will be shown presently, is believed most nearly to represent the original composition of the dike magma.

TABLE 3

	A	*B*	*C*
SiO_2	42.26	40.70	41.46
TiO_2	3.18	3.86	5.70
Al_2O_3	14.61	16.02	14.24
Fe_2O_3	5.38	5.43	3.32
FeO	4.56	7.84	5.70
MnO	.05	.16	.08
MgO	6.12	5.43	13.68
CaO	9.40	9.36	11.62
Na_2O	2.64	3.23	2.29
K_2O	1.48	1.76	1.72
H_2O^-	2.72	2.62	—
H_2O^+	3.93		.12
CO_2	2.99	2.97	—
P_2O_5	.62	.62	—
S	.09	nd	nd
BaO	.16	nd	nd
F	nd	nd	.42
	100.19		100.35
		Less O=F	.18
			100.17
		(Sp. Gravity=3.215)	

Norm (from Column *A*):

orthoclase	8.90
albite	22.01
anorthite	23.35
nephelite	0.28
diopside	15.55
olivine	5.74
magnetite	5.34
ilmenite	6.08
hematite	1.76
apatite	1.34

C.I.P.W. classification:
Camptonose, (II)III.5.3″.4.

A—Camptonite dike from eight miles south of Boulder Dam, F. A. Gonyer, *analyst.*
B—Daly's (1933) average for 15 camptonites.
C—Amiphibole phenocryst from Boulder Dam camptonite, F. A. Gonyer, *analyst.*

Calculation of the norm of this rock presents some difficulties with respect to the $CaCO_3$ content. Washington (1917, p. 1167) states (rule 3*h*) that "if the modal calcite is primary, this normative molecule is to be regarded as femic; if the modal calcite is secondary or due to the assimilation of limestone, the calculated calcite molecule is to be disregarded as not forming part of the norm." In short, one is faced with the necessity of deciding whether the calcite is primary or secondary. Primary calcite is viewed with deep suspicion by some petrographers: Johannsen (1931, p. 150); others appear to accept it without question: Poitevin and Cooke (1946). In this occurrence it does not appear to be an alteration product, nor do we have any real evidence that it has formed as the result of assimilation of limestone. It is perhaps worth noting Pirrson and Washington's (1906, p. 499) remarks on a camptonite from the Belknap Mountains, New Hampshire, in which calcite is present in much the same relationships as in the present rock. They say "calcite . . . does not appear as an ordinary alteration product . . . but rather as infiltrated material, if it is not indeed an original component." In obtaining the norm for this rock these authors (*idem*, p. 501) compromise by calculating in CaO and disregarding CO_2, a practice rather generally followed (Washington, 1917). The present calculation has followed this procedure with the results as given in Table 2. If calcite had been calculated as femic, nephelite and olivine would disappear from the norm, being replaced by quartz (1.51%) and hypersthene (14.90%). This however does not change the classification of the rock as Camptonose.

PETROGENESIS

The source of this camptonite magma is unknown. Its field association with olivine basalt suggests that it might be a differentiate from such magma. However, considerations as to the origin of the magma can be nothing more than speculative with so little known of the general geological history of this region. Since the occurrence lies in a block probably down-faulted against the Black Mountains, it is not unlikely that Upper Paleozoic limestones are present in the section at depth, and assimilation of such rock may be in part responsible for the development of the camptonite magma. To those who may regard camptonites as belonging to a diaschistic series, as first suggested by Brögger (1894), there will be little comfort in knowing that no complementary leucophyres have been discovered in the region.

Of more immediate interest than the origin of the magma is the problem of the amphibole phenocrysts. The typical amphibole of camptonites (Holmes, 1928) ordinarily is barkevikite, which differs from the amphibole of these dikes chiefly in its much higher ratio of magnesia to ferrous iron and in its lower titania and somewhat higher silica content. It is the distribution of the amphibole phenocrysts, however, rather than their composition, that presents the major problem. No one viewing the road-cut exposure can avoid the impression that these phenocrysts must have formed very nearly in place, and consequently under no more than a few hundred feet of cover. If the amphibole phenocrysts were of intratelluric origin, a more or less even distribution throughout the dike would be expected. But as shown previously, these phenocrysts are practically absent from the selvages of this dike and show considerable diversity in development within the main body of the dike. An hypothesis of multiple intrusion might explain this textural diversity. Haff (1939) in his study of the Cape Neddick dikes has illustrated the very great complexity that may develop in such types of intrusion. But none of the criteria of multiple intrusion is recognizable at this exposure, although in certain of the more northerly dikes there are faint suggestions of internal chilled contacts. Multiple intrusion, though it might account for diversity of texture, fails to explain the symmetrical size distribution (see Table 1), unless one makes the unlikely assumption of a series of closely spaced intrusions carrying phenocrysts of progressively larger dimension and each intrusion perfectly bisecting the preceding intrusion! Composite intrusion is likewise an unlikely hypothesis in view of the uniform character of the groundmass throughout the dike.

Before attempting an explanation of the distribution of the amphibole phenocrysts in the rock, certain significant features should be recalled. (1) The olivine phenocrysts, presumably one of the earliest ferromagnesian minerals to form, are found evenly distributed throughout the dike in selvages and central portions alike. (2) All of the olivine has undergone a considerable degree of serpentinization; but the olivine in the chilled margins shows less alteration than that in the central portions of the dike. (3) The amygdules show the same size distribution as do the amphibole phenocrysts, that is, they are larger and more abundant toward the central portions of the dike. Unlike the amphibole phenocrysts they are not absent from the selvages, but occur therein as microscopic vesicles.

Without wanting to put precise limits on the definition of the term "deuteric" (Shand, 1944) is it not significant that both the serpentinization of olivine (Hess, 1935) and the formation of some amygdaloidal minerals (Moehlman, 1936) have been regarded as falling within the category of this process; furthermore, the presence of volatiles is one of the generally recognized conditions for the formation of amphibole (Kennedy, 1935). In the present occurrence, each of these features is clearly related to position within the dike. It seems not unreasonable to suppose, therefore, that the initial intrusion consisted of camptonite magma in which only olivine and magnetite had already crystallized. Volatiles in the magma would find easy escape through the permeable fanglomerate, so easy indeed that no trace remains of their passage. Magma in immediate contact with fanglomerate walls and thus depleted of its volatiles would chill rapidly, thereby providing a fine-grained, dense seal which prevented loss of volatiles to the walls from the remainder of the intrusion. The slowest cooling, more fluid central portions subsequently may have been enriched by access of volatiles from depth. It was the volatiles thus concentrated in the central portion, which were chiefly responsible for formation of the amygdules, which by lowering viscosity permitted growth of exceptionally large amphibole crystals, and which caused serpentinization of olivine. But these volatiles, largely excluded or escaped from the chilled margins, formed practically no amphibole, developed only microscopic vesicles, and produced a lesser degree of serpentinization of olivine in the selvage of the dike. Within intermediate portions of the dike these processes operated to intermediate degrees.

A process, analogous to this in some respects, has been described by Fuller (1938) in which deuteric alteration of columnar lavas has occurred principally in central portions of columns and flows, whereas unaltered rock occurs at the surface of flows and adjoining the vertical fissures on edges of columns. In brief, where volatiles have opportunity to escape, deuteric effects are at a minimum; where gases have been sealed into the rock, such effects have tended to a maximum.

The calcite- and analcime-filled amygdules need little additional comment. Presence of amygdules in intrusive bodies is uncommon, but by no means unknown: Morris (1930); Moehlman (1936). Morris has noted that in most such occurrences gases responsible for formation of vesicles in intrusives have been derived from external sources, such as coal beds and other rocks rich in volatiles. In the present instance, two such possibilities may be suggested: (1) water in the porous fanglomerate —but this is unlikely on the presumption that the fanglomerate is little more than a veneer, and the dike magma had therefore relatively little contact with it; (2) carbon dioxide from assimilation of limestones deeper in the section—but this is no more than speculation. It seems best therefore, to consider, as Moehlman did, that the volatiles were probably of magmatic origin.

To what extent these observations and deductions apply to the other camptonite dikes in the area is not known, for in no other dikes are comparable cross sections exposed. Occasionally in the more northerly dikes small amphibole phenocrysts are found in the chilled margins. In a majority of these dikes the size range shown by the amphibole phenocrysts is not as great nor are the crystals of as large size, although the distribution is of the same type (larger sizes toward the center). These differences may be explained by the difference in level of the exposures: the northern exposures being 300 to 400 feet above the section exposed in the road-cut. Wentworth and Jones (1940, Fig. 8) have shown graphically how rapidly textural characters of dikes may vary at upper levels. This is especially true if, as appears here to be the case, these levels were near enough to the top of the section so that volatiles could escape to the surface through the dike magma, thereby conditioning the development of only smaller phenocrysts and amygdules.

Other puzzling features of the dikes deserve mention. (1) Plagioclase microlites show a high degree of parallelism (flowage texture), amygdules show moderate stretching, but the amphibole phenocrysts show very little dimensional orientation. Why should there be this contrast within the same body? It is believed that the difference in grain size may afford an explanation: the small plagioclase microlites could have been aligned by movements of magma insufficient to orient the very much larger amphibole phenocrysts. The amygdules, of intermediate size, would exhibit intermediate response. (2) Why do the amphibole phenocrysts show a notable size distribution with respect to dike walls, while other minerals do not? Winkler (1949) points out, in his study of the influence of cooling rates on crystal dimensions, that hardly any change in crystal size is to be expected in thin dikes (that is, those with a width of six feet or less). These camptonite dikes, with widths averaging only four feet, should show no change in grain size—which is indeed the case for most of their minerals. It is the amphibole phenocrysts that are aberrant, and to explain them one must assume that extremely favorable conditions of composition (Kennedy, 1935) and of ionic mobility of the requisite components persisted for a sufficient length of time in the central portions of the dikes, with a gradual reduction in these factors toward the margins. Slight difference in concentration of volatiles might produce considerable difference in crystal size, for as Buerger (1948) has recently shown, even small amounts of (OH) and F are structurally very significant in reducing viscosity of magma and in speeding crystal growth. (3) How long did it take the largest amphibole phenocrysts to grow? There is no direct answer to this, but rough computations, following Lovering's (1935) discussion of cooling in a dike, can be made that will indicate how long favorable temperatures may have obtained in the central portion of the dike. For this certain assumptions are necessary. The temperature at which the amphibole began to crystallize was probably not over 800° C., for above this point the mineral might be expected to show the very low extinction angles characteristic of the more highly heated oxyhornblendes (Barnes, 1930). A lower temperature limit of 200° C., based on the analcime filling, is suggested by the formation of amygdules which require a relatively high viscosity in the magma—a viscosity which would probably preclude further growth of the amphibole. 800° C. to 200° C. is presumably an extreme temperature range; in all probability it was far less than this, as suggested by the lack of zoning in the amphibole phenocrysts. No diffusivity constant for camptonite is available, so the value ($h^2 = .0083$) given for basalt by Birch (1942, p. 253) was employed in the calculation. Using these data, a period of about 25 days is obtained for the fall of temperature at the center of the dike from 800° C. to 200° C.

This of course assumes that all heat loss was by conduction through the walls. Actually there may have been loss of heat by gas transport, and—remembering that these are shallow structures—by escape of heat to the surface. Both of these factors would further reduce the 25-day figure. Opposed to this is the strong possibility of transfer of heat by volatiles from below, which might partly balance or even exceed conduction and other losses. At any rate, it remains an interesting speculation that these exceptionally large amphiboles may have grown in a relatively short time. Some have suggested that, considering the anomalies in this occurrence, the amphiboles may best be explained as porphyroblasts rather than phenocrysts, and Smith (1946) in a general discussion of the lamprophyre problem has emphasized that not all the crystallizations therein should be regarded as orthomagmatic. Nevertheless we regard this hypothesis as untenable for these amphiboles. That the metasomatism required to produce crystals of this composition and of this size could operate at the low temperatures and pressures that the general geologic setting requires—much less, that it would leave no mark on the country rock—is unthinkable. The amphibole is igneous.

Mechanics of Intrusion

Wherever well-exposed relatively recent dikes are found, there springs the hope of discovering new data on the mechanics of dike intrusion. In this instance, the rudely defined or almost absent stratification in the fanglomerate country rock precludes any possibility of making accurate measurements of any displacement of walls of dikes, direction of movement, etc. The absence of even local crumpling or contortion in strata adjoining the dikes, as well as absence of any sizable protrusions of the magma into the relatively permeable fanglomerate, suggests that the magma came in under no great pressure. On the other hand, it seems unlikely that an open fissure in fanglomerate could have existed into which magma up-welled. The most plausible assumption is that the dikes formed in fractures, actual or potential, which were widened in part by tensional forces, as magma pushed in. No physiographic or stratigraphic evidence has been observed that would indicate control of the dike loci by faults; but it is significant that the trend of the dikes parallels the "grain" of the area and of the general northerly trend of the major faults of the region (Longwell, 1936).

Age and Correlation

Except for these camptonites, rocks of alkaline affinities have hitherto been unknown in Western Arizona. Indeed, the nearest alkaline rocks are the small bodies of nepheline-syenite described by McAllister (1940) from the Death Valley region, California, over 150 miles distant. The closest approaches to lithologic similarity are the occurrences of camptonite dikes and sills in the Northern Argus Range and the Darwin District, California, reported by Hopper (1947) which he correlates with lamprophyres of the Searles Lake (California) quadrangle described by Hulin (1934) and shown by Hulin to be pre-Middle Miocene and probably early Eocene in age.

Since no fossils have been found in the fanglomerate country rock of these Arizona dikes, the age of the formation is unknown. About a mile south of the mapped area, the fanglomerate lies with angular unconformity upon beds of the Muddy Creek formation (Longwell, 1936) of Lower Pliocene (?) age. This establishes the dikes as post-Lower Pliocene (?). The upper age limit could be approximated only by detailed physiographic studies which are impractical until accurate topographic maps or airphotos of the general area become available. Sklar (1938) has suggested that lavas, with which the flows of the map-area may be correlated, are Pleistocene and possibly Recent. The best conclusion for the time being seems to be that the fanglomerate may be Pleistocene and the Camptonite intrusions late Pleistocene or Sub-Recent.

Summary

This occurrence of camptonite dikes near Boulder Dam is of interest because it provides a convincing demonstration of the control by volatiles of certain crystallization processes; because of the development of a somewhat unusual type of amphibole; because it is the only known occurrence of rocks of this character in Western Arizona; and finally, perhaps, because of the striking exposures and excellent specimens provided on a route frequently travelled by geologists.

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We first visited this locality in the company of Dr. J. Volney Lewis and Mr. Robert H. Rose of the National Park Service, and it was at their suggestion and with their encouragement that the study of these dikes was undertaken. The late Harry Berman provided helpful discussion of the amphibole analysis. Colleagues in the National Park Service and at the California Institute of Technology have aided us in the field and in the laboratory. We have enjoyed stimulating discussions of the curious features of these dikes with many geologists, among whom may be mentioned Dr. V. P. Gianella, Mr. E. D. McKee, Dr. W. E. Powers, Dr. Heinrich Ries, Dr. A. F. Rogers, Dr. E. D. Wilson, and the late Dr. John E. Wolff. To all these persons we wish to express our thanks for aiding this study, whether tangibly or intangibly. To none of them

do we assign any responsibility for the conclusions reached or the hypotheses presented.

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QUARTZ VEINS FORMED BY METAMORPHIC DIFFERENTIATION OF ALUMINOUS SCHISTS

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Abstract

Quartz veins in staurolite schist are believed to have formed by metamorphic differentiation. As silica was removed from the wall rock to form the veins, micas, staurolite, and accessories were concentrated in the schist bordering the veins. The inner part of this border zone is composed of staurolite prophyroblasts arranged in a thin layer against the quartz veins. Out beyond the staurolite layer is a wider layer of quartz-poor schist composed almost entirely of mica but devoid of staurolite. The differentiation is believed to have operated according to the concretion and solution principles but differential compression, due principally to folding, is considered to have played an important role. Vein formation continued over an extended period but the veins ceased growing just before the staurolite layer formed.

Introduction

The writer was first impressed by the occasional discovery of numerous large quartz veins in schist while mapping the bedrock geology in western New Hampshire in 1937. The most striking feature about these veins appeared to be the concentration of large staurolite crystals in the schist along the vein walls. Zones, one to two inches wide and commonly composed almost entirely of staurolite crystals apparently identical with the porphyroblasts of the enclosing schist, were noted in several localities. It was concluded (Chapman, 1946) that the veins influenced the localization of staurolite in the schist walls by transporting components essential to the formation of this mineral. More recent studies show that the problem is much more complicated than was formerly thought. Field work has shown that these veins, with associated staurolite zones, are not rare but are widely distributed through the staurolite schists of the region.

The most detailed studies were made of the veins in the Claremont quadrangle in western New Hampshire (Figure 1). The best examples noted are on the west slope of Croydon Mountain (Blue Mountain) within the boundaries of the Blue Mountain Forest Association (formerly called Corbin Park). Persons unable to obtain permission to enter this private game reserve may see very fine exmaples of these veins along the road just west of Perry Mountain in the town of Charlestown.

It is hoped that this paper will further stimulate interest in the study of analogous veins in other regions and in other rock types that such studies may result in a clearer understanding of the problems involving transfer of material during metamorphism.

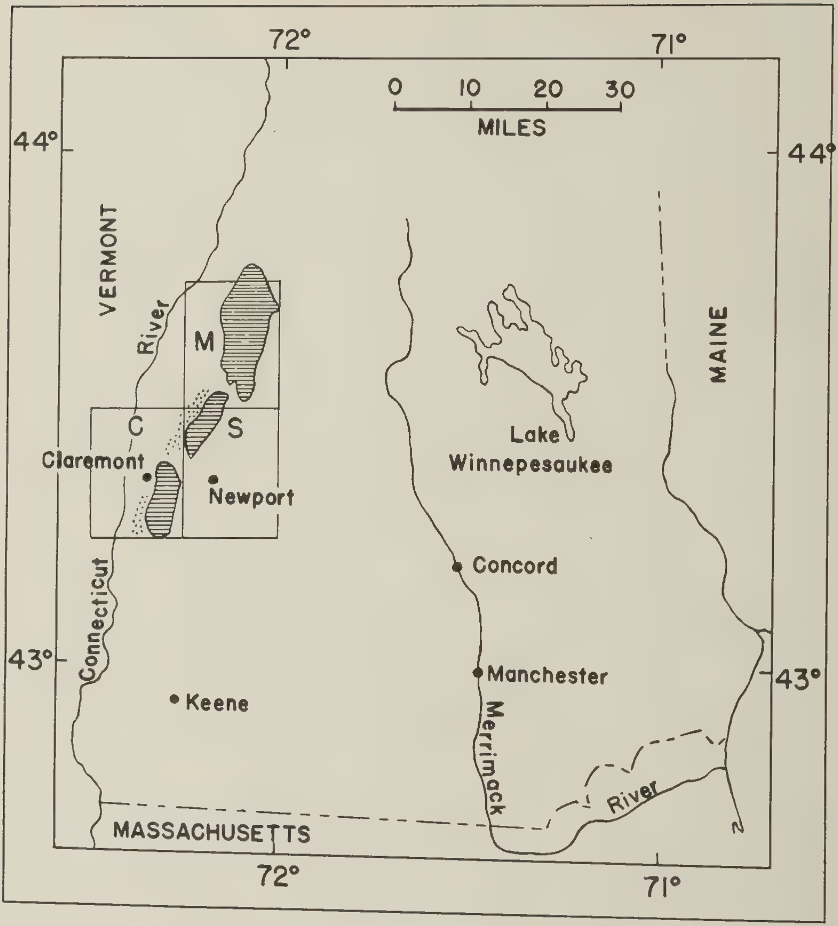

FIG. 1. Index map of southern New Hampshire. Areas where quartz veins were studied—stippled. Domes of the Oliverian magma series—ruled. Claremont quadrangle—C. Mascoma quadrangle—M. Sunapee quadrangle—S.

FIELD RELATIONS AND PETROGRAPHY

General Statement. The rocks considered in this paper occur within a belt of schist which constitutes part of the Littleton formation (lower Devonian) of west-central New Hampshire. This belt lies within the middle-grade zone of metamorphism and extends roughly NNE–SSW along the west flanks of the numerous domes of the Oliverian magma series (Chapman 1939 and 1942). See Fig. 1. The rocks consist chiefly of fine- to medium-grained mica schists, most of which contain prophyroblasts of biotite, garnet, and staurolite; and are believed to have been derived from shale, sandy shale and impure sandstone. A marked schistosity and generally pronounced lineation are characteristicallydeveloped. The general distribution and relative abundance of rock types within the formation are well understood and the detailed petrography has already been worked out (Chapman 1939 and 1942).

Petrography of the Schist. For the present study only the following rock types need be considered; mica schist, quartz-mica schist, and staurolite-mica schist. Staurolite-mica schist is perhaps the most abundant, but it should be noted that these divisions are purely arbitrary and all three types are gradational. For convenience, these schists will be referred to as representing the normal schist type.

In each of the three rock types quartz and plagioclase (oligoclase) occur as elongate or polygonal-shaped grains averaging 0.1 millimeter across. Mica and chlorite may either be scattered through the quartz-feldspar mosaic or concentrated in thin layers. Most of the muscovite occurs as flakes 0.1 to 0.3 millimeters long and is generally oriented parallel to the foliation of the rock. Red-brown biotite is found most commonly as porphyroblasts one to two millimeters across and oriented at various angles to the schistosity. Garnet is not an abundant constituent. It occurs somewhat sporadically as anhedral to subhedral grains up to one or two millimeters across and incloses quartz poikiloblastically. Staurolite occurs as porphyroblasts, up to four or five inches long, which inclose quartz grains or small patches of the groundmass. In most of the staurolitic rocks these porphyroblasts average one to two inches long and locally may be confined to certain layers. Where this difference in stauorlite content is obviously related to bedding, it is believed to be due to differences in chemical composition of the original sedimentary rocks. For the most part the individual porphyroblasts have been completely altered to a fine aggregate composed mostly of sericite with some chlorite. Where alteration has not been so intense, only a shell of sericitic material surrounds a relatively fresh core of staurolite. Most of the chlorite in these rocks is pseudomorphic after biotite, whereas minor amounts only have formed from garnet. The minor constituents of the schists include brownish green tourmaline, zircon, apatite, sphene, ilmenite, pyrite, and magnetite. In Table 1, modes 1 to 5 inclusive represent several varieties of the schist.

Quartz Veins. Quartz veins are numerous in the schists of the Littleton formation. They range up to about two feet in thickness and many tens of feet in length. On outcrop surface they may appear as simple, clean-cut bands or they may grade into highly irregular and anastomosing vein-complexes. Some are sharp, straight veins whereas others are curved, folded, irregular, lenticular or pod-like. The pinch-and-swell type of structure is not uncommon. For the most part the veins follow the schistose structure of the rock, particularly where schistosity and bedding are parallel. Where bedding is folded and schistosity cuts it, as at the noses of folds, the quartz veins generally wrap around with the bedding. This relationship is well shown in Fig. 2 and its significance will be considered later. Even where bedding and schistosity are parallel a few veins have been observed cutting both structures at 10–25 degrees.

Petrographically the veins are composed of very coarse-grained glassy quartz which in thin section shows strong stain shadows and highly sutured borders. Minor constituents of the veins are, in order of decreasing abundance; feldspar, staurolite and mica. Feldspar is a common constituent and generally occurs in masses up to one inch or so across. It is very commonly abundant near the margins of veins. A two foot vein was observed to contain 10 per cent feldspar in irregular masses about four inches across. Most of these masses were clustered in a small section of the vein about two feet by one foot. In this concentration the feldspar made up 50 per cent of the vein material giving the rock a typical pegmatitic appearance. Feldspar is probably much more abundant than one would at first suspect because careful examination of many veins shows that much feldspar has been removed by weathering at the surface of the vien, leaving epimorphs in the quartz. These small cavities become favorable loci for lichens and are eventually almost completely obscured. In thin section some of the more granular aggregates of feldspar show a medium- to coarse-grained, granablastic texture. The feldspar is oligoclase and much is slightly sericitized. Albite twinning is fairly conspicuous and occasionally both albite and Carlsbad twinning are seen. A few of the large grains show marked zoning. Isolated crystals of staurolite, unassociated with schist and wholly enclosed by the quartz veins, were not commonly observed. Staurolite within the veins is generally unaltered or at least less altered than that in the adjacent schist. In areas where staurolite crystals show outer zones of alteration products, the zones are much thinner for those crystals enclosed by quartz veins. Most staurolite crystals in the veins are well formed with {110}, {001} and {010} universally present. Several small veins were observed to contain tiny grains of fresh-looking staurolite, and in one vein these grains were scattered in two thin layers about one-half inch from either vein wall. Small grains of staurolite in the viens may be much more common than surface exposures would indicate; because, as in the case of feldspar, close examination shows that many small grains have been partially or completely weathered out of the vein rock. Micas, independent of schist inclusions, appear to be very uncommon in the quartz veins. Flakes of muscovite and some biotite, scattered through the vein quartz, show up clearly in thin section but these are not readily detected in the field.

Border-Zone Rocks. Between the typical quartz vein and the normal schist is a border zone of characteristically different rock. At the inner part of this border zone, adjacent to the quartz vein, is a relatively thin layer of staurolite. Between the staurolite layer and the normal schist the border zone rocks are extremely rich in mica. The four rock types, therefore, form a distinct zonal arrangement. In passing outward from the quartz vein one traverses, first, a thin staurolite layer and then a somewhat thicker layer of mica-rich schist before reaching the normal schist. Some of the one inch viens, for example, have border zones ranging between one-quarter and one-half inch thick. For veins up to one or two feet wide the border zone ranges up to about six inches in thickness.

The staurolite layer is usually one inch or less in thickness and may be somewhat discontinuous along the vein wall. In places only scattered crystals of staurolite occur in contact with a quartz vein whereas but a few inches or a few feet farther along the contact staurolite may be so abundant as to form a compact layer between the quartz vein and adjacent rock. Masses up to many inches across composed almost entirely of staurolite are commonly encountered. Field studies show that in general the larger quartz veins are accompanied by a greater concentration of staurolite than are the smaller veins. Both megascopic and microscopic examination of the material in these staurolite layers from numerous localities was undertaken. The size of the staurolite crystals in the border-zone rocks at any particular locality is ordinarily the same as that of the staurolite porphyroblasts in the adjacent schist. In most localities studied these border-zone crystals range from one-half to two and one-half inches long, and are generally well developed against the quartz vein.

The mica-rich schist of the border zone possesses an unusually brilliant to silky luster and the finer-grained varieties are phyllitic. This rock differs from the normal schist in two other respects. It is generally coarser grained and is relatively richer in nearly all mineral constituents except quartz and staurolite. The mica-rich schist layer is more persistent than the staurolite layer; and where the latter is somewhat discontinuous or absent, the mica-rich schist comes in direct contact with the vein wall. Away from the vein the mica-rich schist is gradational into the normal schist, but in general the layer is several inches wide. As a rule the mica-rich schist layer is thickest where associated with the larger quartz veins and thinnest about the smaller veins. A three inch vein, for example, might have a layer of mica-rich schist one inch thick whereas the layers associated with somewhat larger veins may range up to four inches or more in thickness.

The gradational character of the outer boundary of the border zone is well illustrated by three modes shown in Table 1. Mode 5 is for a specimen taken 12 inches from the vein wall, mode 6 is for a specimen taken four inches from the same vein wall, and mode 7 is for a specimen taken one inch from the same vein wall. This series shows a marked decrease in amount of quartz+feldspar toward the vein and an increase in amount of micas. Modes 8, 9, 10, and 11 are quite typical of the inner part of the mica-rich schist layer, whereas mode 12 is somewhat richer in quartz.

Table 1. Modes of the Normal Schist and Mica-Rich Schist*

	1	2	3	4	5	6	7	8	9	10	11	12
Quartz plus Feldspar	55	56	64	65	74	66	1	3	1	7	2	30
Biotite	15	14	15	16	10	13	32	28	41	52	58	2
Muscovite	12	19	12	9	14	20	64	67	53	39	38	65
Staurolite	17	3	7	8	1	0	0	0	0	0	0	0
Garnet	1	7	1	1	tr.	tr.	1	1	3	tr.	0	1
Tourmaline	tr.	tr.	tr.	tr.	tr.	tr.	0.9	0.5	0.4	0.2	0.3	0.3
Opaques	0.5	0.4	0.2	0.2	0.3	0.5	1.0	1.0	1.0	1.3	1.3	1.0

1. Staurolite-mica schist, normal schist type.
2. Staurolite-mica schist, normal schist type.
3. Staurolite-mica schist, normal schist type.
4. Staurolite-mica schist, normal schist type.
5. Staurolite-mica schist, normal schist type, 12 inches from quartz vein.
6. Quartz-mica schist, four inches from quartz vein.
7. Typical mica-rich schist, one inch from quartz vein.
8. Coarse porphyroblastic mica-rich schist, one inch from quartz vein.
9. Very coarse porphyroblastic mica-rich schist, one inch from quartz vein.
10. Typical mica-rich schist, one-half inch from quartz vein.
11. Typical mica-rich schist, one-half inch from quartz vein.
12. Phyllite from mica-rich schist layer, 2 inches from quartz vein.

* For simplicity, percentages are given to the nearest unit except for tourmaline and opaques which are expressed to nearest tenth. Chlorite (altered from biotite, staurolite and garnet) is assigned to those respective minerals in the table.

Much of the biotite has been partially or completely chloritized, and biotite and chlorite are as large flakes commonly one to two millimeters long. Many of these large flakes are believed of late development, because they appear to cut off, interrupt and penetrate muscovite-rich layers in a fashion which partially obliterates an earlier crinkled structure. A few biotite plates over 12 millimeters across were noted. In general the cross-micas (biotite) are larger and more abundant nearer the quartz veins, the greatest size and abundance being attained in the inner part of the mica-rich schist layers. In the more extreme cases the individual crystals have developed as long blades rather than as spindles as in the normal schists. In some of the coarser varieties of mica-rich schist, these biotite porphyroblasts are extremely abundant as blades two millimeters wide and 15 millimeters long. Accessory minerals are more abundant and in somewhat larger grains than those of the normal schist.

Inclusions. Inclosed by many of the larger quartz veins are numerous masses of staurolite-rich and mica-rich rocks. These masses appear as stringers, lenses and irregular bodies ranging up to a few feet in length and several inches across. For the most part they are composed of schist not unlike that of the mica-rich schist layer of the border zone. The outer portion (that next to the quartz vein) of many of these inclusions is rich in staurolite and corresponds to the staurolite layer of the border zone. In some of the larger masses the central part is composed of the normal schist with a few staurolite porphyroblasts. This central mass is surrounded by a layer of mica-rich schist which in turn is rimmed by a more or less continuous staurolite layer. The petrographic relationships within the inclusions are, therefore, the same as those for the extra-vein rocks. Numerous stringers and elongate clusters of staurolite crystals within quartz veins are commonly observed on outcrop surfaces. These aggregates, however, are usually associated with minor amounts of mica-rich schist, and they probably represent staurolite layers around tabular inclusions of schist which have scarcely been exposed or which have all but been destroyed by erosion.

Genesis and Significance of Vein Complexes

Age of Veins and Border-Zone Rocks. The quartz veins and staurolite layers cannot logically be considered pre-metamorphism in age because the staurolite must be younger or essentially contemporaneous with the porphyroblasts of staurolite in the adjacent normal schist. The vein complexes are probably not post-metamorphism because they show a marked preference to follow bedding rather than schistosity, where these two structures are not parallel. Figure 2 shows that the quartz veins follow folds in bedding. On the nose of folds the schistosity, which is parallel to that of the region, cuts across the bedding at large angles. These folds and probably the veins as well formed before the final schistosity in this locality. If the veins are neither distinctly pre-metamor-

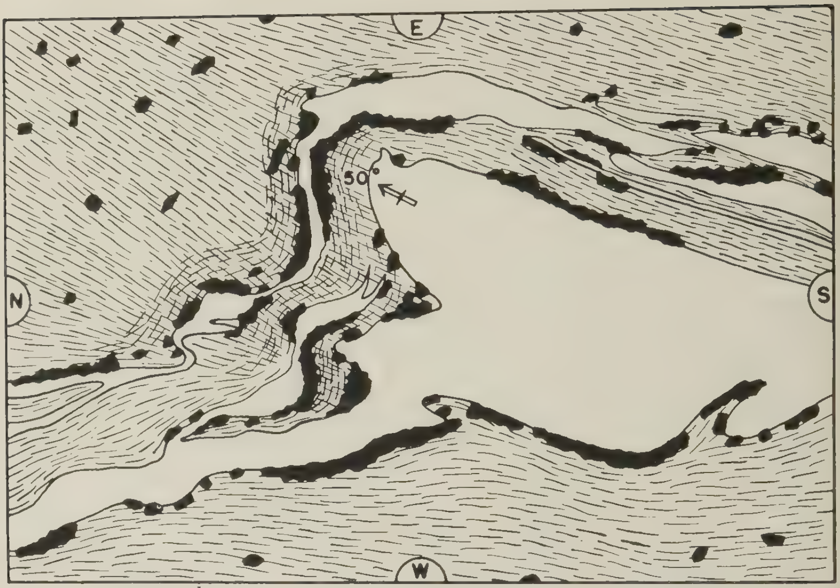

FIG. 2. Map of quartz veins in schist showing staurolite layers. West slope of croydon Mtn. in Croydon, New Hampshire. Area about three feet by four feet. Staurolite—black, quartz veins—clear, schist—lined. Arrow shows attitude and plunge value of folds. Intersecting lines show regional schistosity cutting bedding.

phism nor post-metamorphism in age they may be considered to have formed during metamorphism. Several observations further substantiate this conclusion.

Crinkled structures believed due to relatively late metamorphic processes are much more perfectly formed and better preserved in the mica-rich schist than in the normal schist. This tendency for schistose micaceous rocks or layers to exhibit more perfect crinkled structures than mica-poor rocks or layers is a common observation in metamorphic study. The micaceous portions are weaker and much more sensitive to the forces causing crinkling; and by virtue of their more perfectly developed schistose structure, the micaceous rocks retain and better reveal these small-scaled features. Therefore, the mica-rich schist layer must have existed, in part at least, prior to crinkling.

Microscopic study of certain specimens, however, shows that the minutely crinkled layers of micaceous material in the border-zone rocks (particularly the inner part of the mica-rich schist layer) are more or less obliterated by the great number of large biotite flakes and porphyroblasts which have replaced the rocks. This would indicate, therefore, that some recrystallization (particularly biotite) took place in the border-zone rocks after the crinkles formed. In one vein, near the contact with border-zone rocks (mica-rich schist), biotite occurs as long, slender blades. These blades are all elongated parallel to each other and to the blade-like porphyroblasts of biotite in the adjacent border zone. The mica blades in the vein so closely resemble those of the border-zone rocks that it is concluded that the former represent relicts of the wall rock which were completely inclosed and isolated as the growing vein developed. Some of the quartz veins, therefore, must have continued to grow even after some of the larger porphyroblasts of biotite had formed. These apparently contradicting observations may be reconciled if one pictures the veins and border-zone rocks as forming over a more or less extended period of time during the processes of metamorphism.

Origin of Vein Complexes. It seems logical to exclude an igneous theory of origin for these quartz veins on the grounds that evidence of igneous activity in the vicinity is lacking and the mineral assemblages of the veins and wall rocks are not characteristically hydrothermal. Perhaps even a better reason for excluding this theory is that the data are so compelling in favor of a metamorphic origin for these features. It is unreasonable to assume that these chemically different rock types could have formed through simple recrystallization of originally chemically different sedimentary layers. The existence of such abnormal sedimentary types in such intimate relationship is extremely doubtful. In preference to starting with such intimately mixed and chemically different sedimentary rocks and metamorphosing them into the contrasting rock layers as we now see them, the writer would prefer to start with a more normal and more homogeneous series of shales and sandy shales and obtain the same results through differential movement of material during metamorphism. Evidence which has led the writer to this conclusion, that metamorphic differentiation best explains the quartz vein phenomena, will now be considered.

Perhaps the most convincing evidence that the quartz veins themselves were derived by secretion from a rock of the normal-schist type is the petrographic character of the border-zone rocks. If the veins have formed by extraction of silica from the wall rocks, then we would expect to find an impoverishment in silica in the border-zone rocks and an enrichment in all other constituents. This is exactly what field and laboratory studies show. A brief study of the modes in Table 1 shows that the amount of quartz+feldspar decreases, to practically nothing, as one approaches the vein. Micas, on the other hand, increase proportionally toward the vein where they make up more than 90 per cent of the rock. Even accessories like tourmaline and opaque minerals are relatively more abundant in the inner part of the border zone than in the normal schist. Garnet is as sporadic in the border-zone rocks as it is in the normal schist. The distribution of staurolite is not clearly shown in the table because all border-zone rocks represented there were taken from the mica-rich schist layer where staurolite is virtually absent. Staurolite in the border zone, it will be recalled, is confined to the vein contact in the form of a thin layer where it commonly excludes nearly all other mineral grains. A clearer picture of mineral variation is shown by the generalized curves of Fig. 3. It is apparent that in the mica-rich schist layer the curves for quartz, staurolite, and feldspar are sympathetically related but antipathetic to the curve for mica. Near the boundary of the staurolite layer quartz and feldspar are virtually absent but the staurolite curve rises abruptly as the mica curve drops. An even sharper break is noted at the vein contact where staurolite disappears and quartz makes up most of the vein.

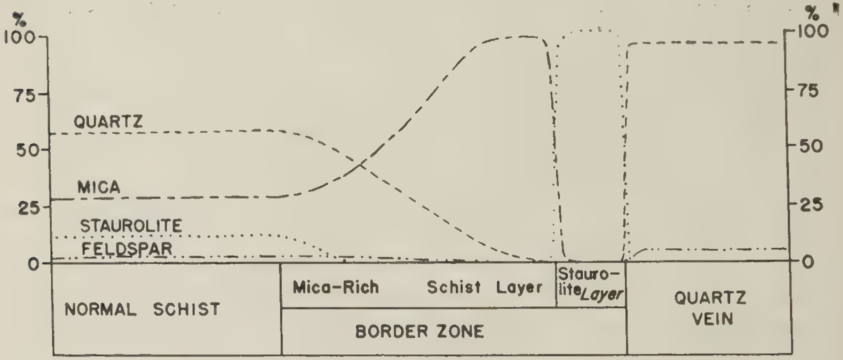

FIG. 3. Variation in mineralogy of the vein complexes.

An attempt was made to check the secretion theory by determining the relative volume of quartz vein, staurolite layer and mica-rich schist layer in a small space and comparing these values with the relative amounts of quartz, staurolite and micas in the normal schist. It was soon realized, however, that a quantitative check could not be very accurate for several reasons. The staurolite layer is somewhat discontinuous along the vein wall. The mica-rich schist is gradational toward the normal schist; and it is difficult, therefore, to set a limit for the thickness of this layer and to determine its average composition. Furthermore, the width of the mica-rich schist layer is somewhat variable. To be accurate, the method would require careful sampling at a very great number of points and it is questionable whether this would be practical.

Quantitative estimates were made and appear to be in harmony with the theory of metamorphic differentiation. Figure 4 is a sketch of a quartz vein complex about three feet by ten inches in which half the area is composed of quartz veins and half of border-zone rocks. No normal schist is shown. The total amount of staurolite in this area is about the same as would appear in an equal area of the adjacent normal schist. Furthermore, the amount of quartz in the veins plus the small amount estimated in the border-zone rocks here totals roughly that of the average normal schist.

Another quantitative estimate may be made by study of Fig. 2. About 200 staurolite crystals occur within the area occupied by the quartz veins and associated border-zone rocks. The average cross-section of these crystals, as exposed, is one-half a square inch. There are exposed, therefore, about 100 square inches of staurolite within an outcrop surface of about 1150 square inches. Staurolite thus constitutes about 8.7 per cent of the outcrop surface. A similar calculation, on a smaller scale, on the normal schist of the vicinity indicates about 10 per cent staurolite. Two similar but large scale calculations in other parts of the formation show 10 per cent and 13 per cent staurolite, respectively. Estimates at numerous localities in the field and even data from thin sections indicate that the staurolite content of these schists is around 10 per cent. It seems reasonable, therefore, that the amount of staurolite (8.7 per cent) in the quartz-vein complex could have segregated during metamorphic differentiation of the normal schist.

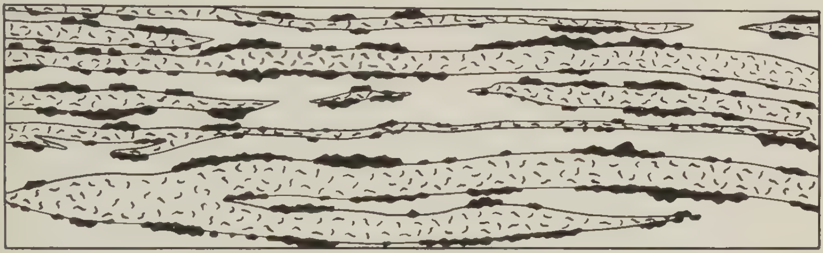

FIG. 4. Map of quartz veins with staurolite layers in mica-rich schist. Near road west of Perry Mtn. in Charlestown, New Hampshire. Area about three feet by ten inches. Staurolite—black, quartz vein—vermiculated, schist—clear.

Origin of the Staurolite Layer. It is not sufficient to say that while the quartz veins grew the staurolitic material was left behind to form the staurolite layers. Several observations must be explained such as why staurolite is so abundant next the quartz veins and why the mica-rich schist is nearly devoid of staurolite. Other minerals besides staurolite were "left behind" but these are not confined to the walls of the quartz veins; staurolite, when present, always intervenes.

The following explanation is offered. The veins started to develop in the earlier part of metamorphism, while the adjacent rocks were being metamorphosed, and continued to grow for some time. Staurolite had not yet formed but staurolitic components as well as components for other minerals (except quartz and feldspar) were becoming enriched in the rocks immediately adjacent to the growing veins. Perhaps micas had already formed; and as more silica moved inward to enlarge the veins, the existing micas near the veins were in part replaced and the components moved outward to the gradually retreating and widening border zones which were becoming richer in those minerals not forming the veins. By the time the veins had grown to nearly their present size, the concentration of staurolitic components in the border zones was great (several times that of the normal schist). When conditions were suitable staurolite started to grow in regions of highest concentration, namely at the walls of the quartz veins. The initial loci of the large crystals were established, therefore, as a result of concentration. The crystals in the staurolite layers appear identical with the staurolite porphyroblasts of the normal schist because they all formed under essentially the same conditions. The higher concentrations of staurolitic material in the border zones resulted in closer spacing of crystals in the staurolite layers. Rapid crystal growth tended to deplete the inner part of the border zones of material to form new crystals and diffusion of staurolitic components toward the vein walls resulted. In effect, staurolitic material was "drawn" for some distance toward the vein contacts and precipitated there as staurolite crystals; this left the mica-rich schist layers devoid of staurolitic material for distances up to, perhaps, a foot from the quartz veins. Beyond this distance staurolitic material was "drawn" outward toward the growing staurolite porphyroblasts of the normal schist. All except the innermost part of the border zone was thus sapped of staurolitic material. The strong power of staurolite to form porphyroblasts explains why nearly all other crystals were excluded from the staurolite layer.

Had not the staurolite grown late in the history of the vein, we would have expected staurolite porphyroblasts scattered widely as inclusions and relicts throughout the vein. Such crystals are rare. Evidence of vein widening against mica-rich schists has already been given. Had the veins grown much after the staurolite layers formed, however, many crystals of staurolite would have been cut off (replaced) by the veins or else segments of staurolite layers would have been surrounded and included in the veins. Such relationships were not observed. Had the staurolite crystals formed early in the history of the quartz veins, they would have been dissolved and re-precipitated time and again until the veins reached their present size. This process would afford no means of cleansing the mica-rich schist layers of the growing porphyroblasts of staurolite. It seems evident, therefore, that the veins received their initial development in the earlier stages of metamorphism, and that they were nearly completely formed before staurolite crystallized.

This theory equally well explains two other observations; (1) why some smaller veins located close to larger veins never developed staurolite layers, and (2) why some small veins developed staurolite layers to an extent disproportionate to their size. These two relationships are illustrated by Fig. 5. Suppose the large vein *A* of Fig. 5 is to grow to form a staurolite layer on either side and that the staurolitic material is drawn from the region between the two dashed lines. The dashed lines represent the outer limits from which staurolitic material moves toward the vein. Suppose a second vein *B* starts to form after vein *A* has attained nearly its maximum size. Before this smaller vein can build up a fair concentration of staurolitic material near its borders, this material will be robbed by the growing staurolite layer of the large vein. Vein *B* will be left without a characteristic staurolite layer. Vein *C*, on the other hand, might develop sufficiently early to build up a rich concentration of staurolitic material near its borders where crystallization would start before staurolite forms at the walls of vein *A*. Once started, the staurolite layer border-

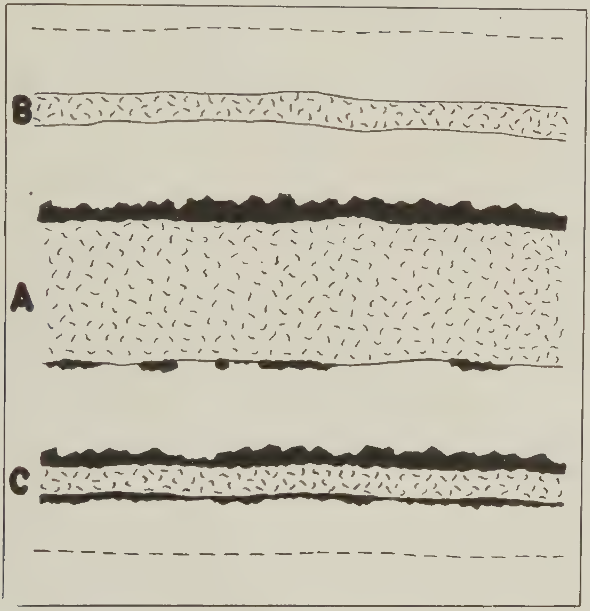

FIG. 5. Idealized sketch showing inconsistencies in thickness of veins and associated staurolite layers. Staurolite—black, quartz vein—vermiculated, schist—clear.

ing vein *C* could continue to draw additional material, made available by formation of vein *A*, from the zone within the dashed line. Vein *A* is thus robbed of much staurolite, whereas vein *C* develops a staurolite zone out of proportion to its thickness. Some veins may grow largely from one side so the staurolite layers become thicker on one side of the vein than on the other. This may account for certain large veins with little or no staurolite on one side whereas the next adjacent vein possesses a normal amount. This gives but three examples of the complexity of the problem. The reader will undoubtedly see others, too numerous to consider here. These three examples are sufficient to explain why some discontinuous staurolite layers may form, why the thickness of a staurolite layer may vary along a uniformly thick vein, and why the thickness of a staurolite layer may be out of proportion to the size of its associated vein.

So far the writer has tried to simplify the problem by considering movement of material essentially perpendicular to vein walls. When it is realized that material must have migrated for some distance parallel to vein walls as well, even greater variations are expected. Some variation in amount of staurolite along veins must, of course, be attributed to irregular distribution of staurolite components in the original sedimentary rocks.

Origin of Inclusions. The inclusions and septum-like masses of mica-rich schist rimmed by layers of staurolite and enclosed by quartz veins are considered to have originated in essentially the same manner as the border-zone rocks. Many of these sheets of included material probably formed while two or more parallel veins widened themselves by addition of silica from the gradually dwindling, intervening layers of schist. They represent, therefore, relicts of the country rock deprived of its quartz and are so differentiated that each possesses its own staurolite layer. This is further substantiated by the observation that the schistosity and lineation of the isolated masses are respectively parallel to the schistosity and lineation of the nearby schist. Furthermore, there are all gradations between inclusions and well defined septa or inter-vein layers.

Longitudinal Migration. The problem of longitudinal migration (parallel to vein walls) of mineral components can not be ignored. Not all irregularities and inconsistencies in the staurolite and mica-rich schist layers can be attributed to such factors as original composition and interference of one vein by another. Some of the more radical variations indicate that longitudinal migration of mineral components took place.

One isolated quartz vein, one to three inches wide and exposed for 15 feet, cuts sharply across bedding and schistosity. The vein strikes N 15 W and dips 75 degrees SW; the bedding and schistosity strike N-S and dip 30 degrees E. In the north half of the vein staurolite crystals form continuous layers along either vein wall. These layers range up to two inches thick. To the south the vein contacts are sharp and not much staurolite is found near the walls. A little further to the south staurolite is again found concentrated at the vein contacts. There appears to be no relation between the amount of staurolite along the vein walls and the staurolite content of the different beds transected by the vein. The staurolite layers are continuous and quite uniform in thickness regardless of whether they cross staurolite-rich or staurolite-poor beds. This strongly suggests longitudinal migration for distances of several feet at least. Furthermore, the same vein shows fresh grains of staurolite (average one millimeter across) distributed in two thin layers parallel to vein walls and located about one-half inch from either wall. These relationships indicate that the staurolite was deposited as a thin layer on either wall of the growing vein. The vein is, without doubt, of the filled-fissure type and indicates longitudinal migration of perhaps many yards.

At another locality several veins up to one inch appear to fill fissures parallel to schistosity. They pinch and swell slightly but do not branch or die out suddenly as do most of the veins. It was noted, furthermore, that they are associated with neither a staurolite layer nor a mica-rich schist layer; and yet, if traced several yards along the strike their thickness increases to two inches and thin border zones are apparent. From here they may be traced to the larger quartz veins with typical border zones. The thinner portions of these veins carry numerous staurolite grains and coarse flakes of biotite and muscovite a few millimeters across. Tiny epimorphs, probably after feldspar, are present. These relationships indicate that some of the segregation veins gradually pass into filled-fissure veins whose components were transported longitudinally.

The occurrence of feldspar in large masses comprising 50 per cent of a vein within an area two feet by one foot has already been described. Such a concentration without longitudinal migration seems very unreasonable in view of the low feldspar content of the normal schist.

Mica-rich schist layers, whose thicknesses are inconsistent with those of the adjacent quartz veins, have been described. Where these occur on both sides of a vein, longitudinal migration of silica may have played an important role. Where the mica-rich schist layer is abnormally thin, it is probable that the associated quartz vein derived much of its silica by longitudinal migration. Where veins are abnormally thin for their border zones, a large percentage of silica may have been removed by longitudinal migration.

Mechanics of Vein Formation. It seems fitting to inquire further into the cause of quartz vein formation. It is conceivable that the well-known concretion principle (Eskola 1932b) could have been operative during metamorphism of the Littleton formation thereby causing the more siliceous layers to become still richer in silica and the alumina-rich layers to be still further enriched in alumina. Such a process would in general promote the formation of relatively thin but uniform quartz layers or veins alternating with micaceous layers. Such features, moreover, would generally be widely distributed provided the original rocks were essentially uniform over great areas. Under what conditions were the quartz veins formed, therefore, that they should be so erratically distributed and vary so greatly in thickness and extent? Why should the segregation of quartz be on such a large scale locally and practically inconspicuous in much of the adjacent schist?

Another question to be answered is why the quartz should segregate in the first place. It is generally agreed that the degree of mobility (solubility) of silica is high but its power of segregation is low for it rarely shows porphyroblastic form. It would seem, therefore, that the role of silica in the formation of these veins was a passive one; and that silica was probably forced into new positions. The precipitation of quartz at the corners of "eyes," developed around porphyroblasts of garnet, magnetite, and biotite, and the quartz fillings in small tension fractures of stretched and broken crystals and pebbles are but a few examples illustrating the tendency for quartz to seek regions of low pressure. The writer does not mean to imply that the concretion principle is ineffective. It is admitted that metamorphic banding in pelitic schists may be due to a higher segregation power of micas, etc., which causes these minerals to concentrate into layers and exclude the quartz which is forced out form to alternating parallel layers. It does not seem, however, that large scale transfer of these materials perpendicular to bedding, due to high power of segregation (the concretion principle), would take place for distances much exceeding one or two inches; and transfer would very commonly be limited to distances of less than one-half inch. Again it seems some additional factor must be considered to account for the distances involved in transfer of material in the formation of the vein complexes.

The solution principle (Eskola, 1932b) must have been operative. It is suggested that during folding, which accompanied metamorphism, potential openings developed parallel to bedding, particularly at the noses of folds. The most mobile materials (silica) flowed from the flanks of the folds to the crests or troughs where pressure was at a minimum. Accompanying the deformation by folding, fractures must have formed. The more open of these must have been filled with quartz. Limited amounts of feldspathic, staurolitic, and micaceous material moved along with the silica into the growing filled-fissure veins. Examples of filled-fissure veins passing into segregation veins have already been given. Squeezing on fold limbs must have been somewhat irregular causing the solution, migration, and deposition of quartz (and some staurolite) along bedding planes to be somewhat uneven. This may partly explain the interrupted staurolite layers as well as the inconsistencies in thickness of the mica-rich schist layers and associated quartz veins.

In summary, it is advocated that, accompanying changes involving the concretion principle, there was differential movement and slipping along bedding surfaces during folding with the result that the more mobile (soluble) substances (quartz and feldspar) were transferred from points of high pressure (on fold limbs) to points of low pressure (at crests and troughs and along tension joints) thereby impoverishing certain layers of the schist in these mobile substances and enriching these layers in less mobile constituents. The quartz, for example, was filter-pressed out of the schist leaving a quartz-poor or mica-rich rock.

Comparison with Other Areas. Read (1933) has made an excellent study of somewhat analogous vein complexes in the metamorphic rocks of Unst, Shetland Islands. Here are quartz-kyanite veins which he attributes to metamorphic differentiation. The features of these quartz-kyanite rocks differ slightly from those of the quartz-staurolite rocks from New Hampshire.

In the first place, Read found kyanite intergrown with quartz along vein walls as well as concentrated in what the present writer would term the border zone. Moreover, kyanite was an important constituent of some of the simple veins. In the New Hampshire rocks, however, staurolite is generally absent from the quartz veins and confined to the inner parts of the border zones.

In the second place, Read considered the vein formation due to lateral secretion, but it is not clear how much, if any, longitudinal migration has taken place during vein growth. In the New Hampshire veins, it is believed that some longitudinal migration was essential.

In the third place, Read considered the concentration of kyanite at the borders of the veins due to impoverishment of the marginal zones in silica. Whereas this is believed to be largely the case with staurolite in the New Hampshire rocks, it has already been shown that some migration of staurolite toward the vein has taken place. It is apparent from Fig. 3 that staurolitic components have evacuated the mica-rich schist and presumably joined similar components in the staurolite layer. This migration across bedding and foliation amounts to nearly a foot in some cases.

In a classical paper by Eskola (1932*a*) brief consideration is given to staurolite porphyroblasts in mica schist in Karelia. His sketch (Eskola 1932*a*, p. 58) showing two small quartz veins roughly parallel to schistosity with staurolite porphyroblasts concentrated at their margins, might easily pass as one representing veins from the New Hampshire locality. Eskola's description of these veins appears in harmony with the ideas and theory advanced in the present paper.

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PETROGRAPHY OF THE NICKELIFEROUS NORITE OF ST. STEPHEN, NEW BRUNSWICK

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Abstract

The St. Stephen complex comprises harzburgite, olivine-norites (including banded troctolitic and anorthositic types) and olivine-hypersthene gabbro, of presumed Devonian age. It is intruded into quartz-chlorite-mica phyllites of the Charlotte Group, which show little contact metamorphism. In contact with the phyllites, the complex has a marginal zone of quartz-hornblende-gabbro. The norites enclose xenoliths of free-silica hornfels (including biotite, cordierite and hypersthene types), silica-poor hornfels (plagioclase-spinel) and pyroxene-granulite. Except in the marginal zone and near xenoliths, hydroxyl-bearing minerals are relatively scarce. Pegmatitic segregations in the norite carry concentrations of ilmenite and show a higher soda/lime ratio than the norites, but no increase in potash, which is low throughout the complex. Lenticular bodies with magnetite, pyrrhotite, pentlandite and chalcopyrite were formed by metasomatism of the norite, associated with development of hydroxyl-bearing minerals which were, however, themselves destroyed where the sulphide mineralization is most intense. A post-sulphide biotite-spessartite dike is recorded, and the development of anthophyllite and chrysotile along post-sulphide fractures is noted.

Introduction

The border town of St. Stephen, in Charlotte County, New Brunswick, is situated on the St. Croix river, 80 miles S.S.W. of Fredericton, opposite the town of Calais, Maine. At St. Stephen and in the district north and west of the town, a complex of basic intrusives covering some nine square miles is exposed, containing concentrations of pentlandite-bearing pyrrhotite. During 1932–35, while I was working in Professor Larsen's laboratory at Harvard as a Commonwealth Fund Fellow, I paid several visits to St. Stephen to carry out the field work which forms the basis of the present study.

It is a pleasure to be able to pay tribute here to the inspiration and guidance Professor Larsen has given to several generations of petrographers, and to mention the esteem in which his work is held, far beyond the confines of his own country.

The laboratory work has been continued in England. It had been hoped that circumstances would make it possible to resume and amplify the field work. As, however, the war and postwar situations have made it unlikely that this hope will be fulfilled, it has seemed best to record the results so far obtained.

The cost of the chemical analyses has been defrayed by a grant made in 1939 by the Government Grants Committee of the Royal Society. I also wish to acknowledge my gratitude to Professor L. C. Graton for opportunities to examine polished sections made by his method at

Harvard, and to Dr. J. Phemister for laboratory facilities at the Geological Museum, London, during the earlier stages of the work; I also wish to record my thanks to Mr. and Mrs. O. W. Dunham of St. Stephen, my hosts in the field.

HISTORY OF RESEARCH

The first description of the area is due to L. W. Bailey and G. F. Matthews (1872, p. 32), who remark:

> "Just north of the town of St. Stephen, and occupying a position between the syenitic gneisses . . . and the mica schists which appear to overlie them, are limited outcrops of dioritic rock which differ in many respects from any strata that we have met elsewhere. They are of very dark, almost black colour, with a very evident and regular stratification, often including layers of dark green serpentine, while in the mass of the rock, besides white feldspar and hornblende—the latter predominating—crystalline masses of diallage are occasionally met with. These rocks, which have been examined *in situ* by Dr. (T. Sterry) Hunt, and by whom the serpentines have been found to yield both chrome and nickel, are apparently connected with the underlying syenitic gneiss rather than the schistose rocks which succeed them to the northward."

H. P. H. Brummell (1892, p. 112) records prospecting operations for nickel and copper in "coarse-grained diorite" on the Todd, Ganong and Carroll properties. These prospects are again mentioned by R. W. Ells (1903, p. 156) who states that pockety masses of nickel-bearing sulphides occur in gabbros which have penetrated black and grey slates of doubtful age, the slates being metamorphosed locally to chiastolite- and mica-schists. C. W. Dickson (1906, p. 236), who made a petrographical study of material from the Ganong property, showed that the sulphides corrode and replace, along lines of brecciation, an intrusive rock which he regards as transitional between gabbro and diabase. He considered that the sulphides were "prominently associated with hornblende and chlorite," and were therefore not direct magmatic segregations. In a paper among the first to record the application of metallographic technique to the investigation of natural opaque minerals, W. Campbell and C. W. Knight (1907) described the relations of the sulphides in the ores of Sudbury and also those of St. Stephen. In 1929–30 New York interests prospected an area north of the town, according to an account by B. Low (1930) in which results of geophysical investigation, diamond drilling and metallurgical tests are given. In 1946 a preliminary geological map, by G. S. Mackenzie and F. J. Alcock, with descriptive notes, was published by the Geological Survey of Canada.

FIELD RELATIONS

The St. Stephen complex forms a small part of a plutonic body of batholithic dimensions which is mapped along the Atlantic coast of

Maine from Frenchman Bay to Englishman Bay, and which then turns inland through Whitneyville and Grove to Calais (Keith, 1933). In New Brunswick it continues E.N.E., passing north of St. George and extending beyond Bayard and Blaydon. The total length is about 140 miles and the average width about 12 miles. The St. Stephen area is an embayment on the north-west side of this pluton, separated from the main mass on the Canadian side by a tongue of Palaeozoic rocks between St. Stephen and Oak Bay, but linking southwards with the main mass in the United States. The rocks of the pluton are classified as 'St. George Intrusives' by MacKenzie and Alcock (1946) on the Canadian side; they include, in the neighborhood of St. George, pink microcline-bearing granite, and dark gray quartz-augite-diorite which have been quarried for building stone.

The norite-complex occupies the ridges and valleys for a distance of some 2½ miles north and west of the town of St. Stephen. There is a patchy but generally thin cover of glacial drift, but a reasonable number of exposures can be found.

The complex is intruded into a regionally-metamorphosed formation of phyllites, with silty bands and beds of quartzite, the 'Dark Argillite' division of the Charlotte Group, probably of Ordovician age (MacKenzie and Alcock, 1946). This group contains minor intercalations of volcanic rocks, and, beyond the limits of the area examined, it passes into schists and gneisses. Adjacent to the norite, on Old, Basswood and Pomeroy ridges, phyllites striking E.N.E. and dipping steeply to the south are exposed; cleavage appears to be parallel to bedding here. A number of large inclusions derived from these rocks are found within the pluton (Fig. 1); since the strike and dip is the same in these as in the surrounding country, they might be regarded as roof-pendants. The age of the batholith as a whole is regarded as Devonian since it intrudes the Silurian Oak Bay Formation, and has yielded pebbles to the Upper Devonian Perry Formation.

The northernmost exposures of igneous rock on Old Ridge reveal a mass of dark rock which proves to belong to the peridotite family. The remainder of the complex is dominated by olivine-norites of various types but the mass is far from homogeneous. It includes, particularly along Hanson Road and in the small stream valley north of the town, banded norites with alternations of anorthositic and troctolitic types—the 'stratified' rocks noticed by the early investigators. It is notable that the strike of the banding, which is nearly vertical, conforms with the Caledonoid trend of the invaded sedimentary formation (Fig. 1). Anorthositic norite is well developed in the neighborhood of Elm Street, and along Cemetery Road; elsewhere, gray rock of gabbroid appearance

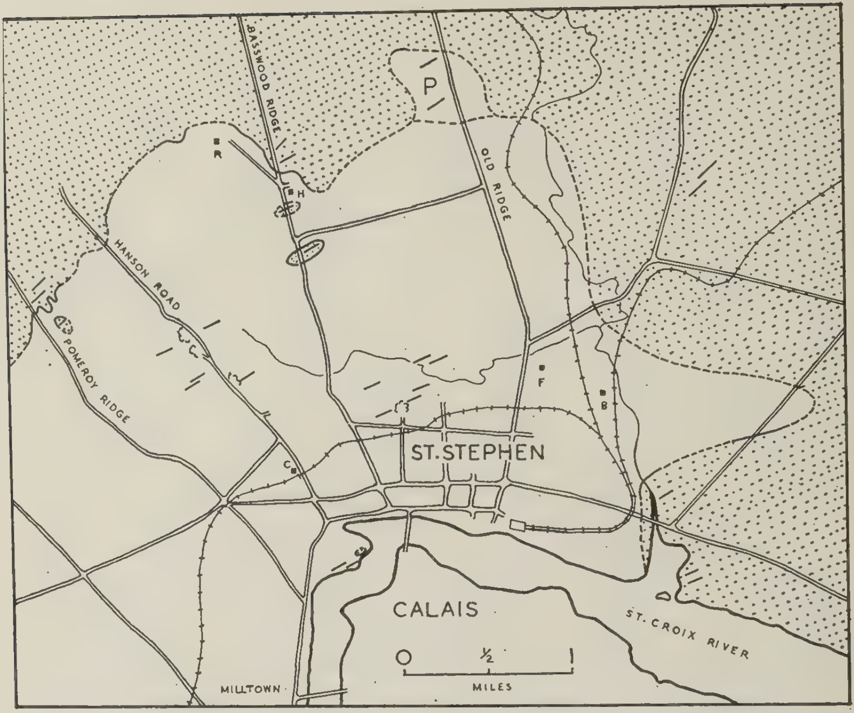

FIG. 1. Geological sketch-map of the St. Stephen district. Stipple—Charlotte Group (pelitic sediments); No ornament—Norite-complex; P, periodotite; R, Rodgers prospect; H, Hall-Carroll prospect; F, Frogtown prospect; B, pit between railroads; C, Cemetery Road excavation. Short black lines indicate strike of banding in norite and bedding in sediments.

prevails. The varieties appear to grade into one another, and exposures are insufficient to justify the mapping of lines between them. The margins of the pluton can be located accurately on Pomeroy, Basswood and Old Ridges, along the road leading north-east from the bridge over Denny Stream, and near the mouth of the stream; the remaining lines are interpolated. Within 100 yards of the margin of the complex, norite gives place to quartz-gabbro rich in hornblende, dark green in color.

Sulphide bodies have been explored at several places. The largest lens so far discovered is on the Rodgers property (R on Fig. 1), west of Basswood Ridge Road, 2½ miles N.N.W. of St. Stephen. According to Low (1930, p. 116) this measures 240 × 112 ft. at surface and has been followed by drilling to a depth of 250 ft.; he estimates the tonnage here at 150,000, containing 45.5 per cent Fe, 24.5 S, 0.9 Cu, 1.5 Ni, 0.1 Co, 0.1 Zn. The long axis of the lens trends E.N.E. On the east side of Basswood Ridge Road occurs a vein-like mass striking north-south on the

Hall-Carroll property; this is 400 ft. long, 10–28 ft. wide (H on Fig. 1). Other occurrences have been tested on the Ganong property, adjacent to the St. Croix river near Milltown, and in pits south-east of Denny Bridge. A showing of sulphides was found in an excavation for a water-main in Cemetery Road in 1935, and magnetic anomalies have been found at numerous other places (*vide* Low 1930, Fig. 2).

South of the area examined, norite gives place to granite. MacKenzie and Alcock (1946) state that the granite cuts a mixed zone of gabbroic and dioritic rocks.

Petrography*

In the account which follows, all determinations of non-opaque minerals are based upon refractive index measurements carried out by the immersion method, which Professor Larsen's work (Larsen, 1921; Larsen and Berman, 1934) has done so much to implant firmly in the technique of petrography.

Harzburgite.—The peridotite exposed on Old Ridge is composed of olivine in rounded crystals up to 2.5 mm. diameter, with subordinate interstitial orthopyroxene and occasional plates of dark brown biotite. Only small amounts of feldspar are present. The composition of the olivine, from refractive index measurements, is Fo_{76}. Closely-spaced veinlets of antigorite with magnetite granules traverse the rock, parallel to the regional E.N.E. strike; these are far more prominent where they cross the olivine than in the pyroxene. In measuring the approximate mode given in Table 1, the chlorite and secondary magnetite were disregarded. The rock is classified as a harzburgite. Other occurrences of peridotite, partly invaded by sulphide mineralization, were noted at the Frogtown prospect (F on Fig. 1) and near a trial pit between the branches of the C.P. railroad.

Olivine-Norite.—The average type of norite is represented by analysis No. 5623 in Table 2 and by a mode in Table 1. The rock, from the St. Croix river cliff, $\frac{1}{2}$-mile west of the railroad station, is a dark gray lustrous phaneride, dominantly composed of a xenomorphic-granular aggregate of plagioclase of composition An_{70-73} (labradorite-bytownite), slightly zoned in places to An_{66} round the margins of the crystals. The average length of the plagioclases is 2.0 mm. Rounded crystals of colorless olivine up to 1.0 mm. diameter are mantled with feebly pleochroic orthopyroxene of composition En_{75} (hypersthene); these also enclose pyroxene crystals. A subordinate amount of clinopyroxene having $n\text{Y} = 1.708$ is

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* A representative selection of rocks and slices from the area has been presented to the Sliced Rock Collection of the Geological Survey and Museum, London; these carry the registered numbers F.4888, F.5596 to 5659.

TABLE 1. MODES OF ST. STEPHEN ROCKS

(Weight—percentages)

GSM No.	*Rock name* Locality	Modal Plagio-clase	Plagioclase	Olivine	Orthopyroxene	Clinopyroxene	Hornblende	Chlorite, bowlingite, talc	Biotite	Clinozoisite	Ores	Quartz	Calcite	Johann-sen Family
5622	*Harzburgite*, Old Ridge	An_{70}	2	77	18	—	—	*	2	—	1	—	—	428
5604	*Troctolitic olivine-norite*, Thompson Pasture quarry	An_{70-75}	50.6	40.5	2.9	—	—	0.5	0.6	0.5	4.4	—	—	?3312P
5646	*Olivine-hypersthene-gabbro*, N. of Hall-Carroll prospect	An_{70-73}	54.6	21.1	7.8	9.6	5.3	—	0.3	—	1.2	—	—	2312P
5623†	*Olivine-norite*, ½-mile W. of C.P.R. station, St. Stephen	An_{70-73}	55.0	20.1	11.1	8.1	0.7	1.3	0.5	—	3.2	—	—	2312P
5607	*Anorthositic olivine-norite*, W. of Elm St.	An_{71}	75.1	9.3	1.4		12.8		—	—	0.9	—	0.5	2312P
5614	*Anorthositic olivine-norite*, Cemetery Rd.	An_{70}	78.4	12.6	1.0		5.8		—	0.2	2.0	—	—	2312P
5605†	*Anorthositic olivine-norite*, N. end, Elm St.	An_{70}	81.8	2.6	2.3		12.3		—	—	0.8	—	—	2312P
5648†	*Norite-pegmatite*, Basswood Ridge, N. of Hall-Carroll prospect	An_{50-32}	66.0	—	—	—	15.5	9.4	—	—	5.2	1.5	2.4	238

* Measured with olivine and clinopyroxene; see p. 715.

† Chemically analysed rock; see Table 2.

TABLE 2. CHEMICAL ANALYSES AND NORMS OF ST. STEPHEN ROCKS

Analyses	5623 Olivine-norite $\frac{1}{2}$-mile W. of C.P.R. station, St. Stephen	5605 Anorthositic olivine-norite, Elm Street, St. Stephen	5648 Norite-pegmatite N. of Hall-Carroll prospect, Basswood Ridge	4888 Mineralized norites, Hall-Carroll prospect	5655 Rodgers prospect
SiO_2	46.14	46.88	48.82	36.26	25.06
Al_2O_3	19.58	28.97	16.93	15.74	7.02
Fe_2O_3	0.64	0.09	0.51	—	—
FeO	8.51	3.01	8.48	—	—
MgO	9.78	3.62	6.08	10.50	4.27
CaO	10.80	14.02	9.42	6.22	2.70
Na_2O	1.58	1.87	3.14	1.58	0.81
K_2O	0.19	0.14	0.15	0.15	0.39
H_2O+	0.65	0.83	0.69	2.84	0.73
H_2O-	0.26	0.27	0.26	1.20	0.20
TiO_2	0.78	0.12	3.39	0.44	0.77
P_2O_5	0.04	tr.	0.48	0.07	tr.
MnO	0.15	tr.	0.24	0.12	0.46
CO_2	tr.	nt.fd.	1.16	nt.fd.	tr.
Fe_7S_8	0.74	—	—	15.03	38.32
$CuFeS_2$	—	—	—	—	1.21
Residual Fe as FeO	—	—	—	8.84	16.70
NiS	nt.fd.	—	—	0.42	1.14
Co	—	—	—	nt. fd.	tr.
Zn	—	—	—	nt. fd.	nt. fd.
	99.84	99.82	99.74	99.41	99.78
Norms					
qu	—	—	1.62	—	—
or	1.11	0.83	0.89	0.89	2.22
ab	13.10	15.77	26.51	13.62	6.81
an	45.87	69.61	31.69	30.30	13.34
cor	—	0.20	—	—	—
di —wo	3.13	—	1.97		
en	1.90 } 6.08	—	1.00 } 3.89		
of	1.05		0.92		
hy—en	8.70 } 13.58	3.63 } 5.54	14.20 } 23.31		
of	4.88	2.01	9.11		
ol —fo	9.73 } 15.85	3.86 } 6.41	—		
fa	6.12	2.55			
mt	0.94	0.14	0.70		
il	1.52	0.20	6.38		
ap	0.10	tr.	1.01		
ca	tr.	—	2.74		
	98.11	98.70	98.74		
pyrrhotite	0.74	—	—	15.03	38.32
water	0.91	1.10	0.95		
	99.76	99.80	99.69		
Normative plagioclase	An_{78}	An_{81}	An_{54}	An_{69}	An_{67}

Analyst, W. H. Herdsman, Glasgow, 1939.

associated with the hypersthene. Rod-like inclusions of an opaque mineral form patches in both pyroxenes. Small quantities of amphibole, pleochroic pale brown to yellow (further discussed below), and of biotite have developed at the expense of the pyroxenes. Scattered crystals of magnetite and of pyrrhotite are present.

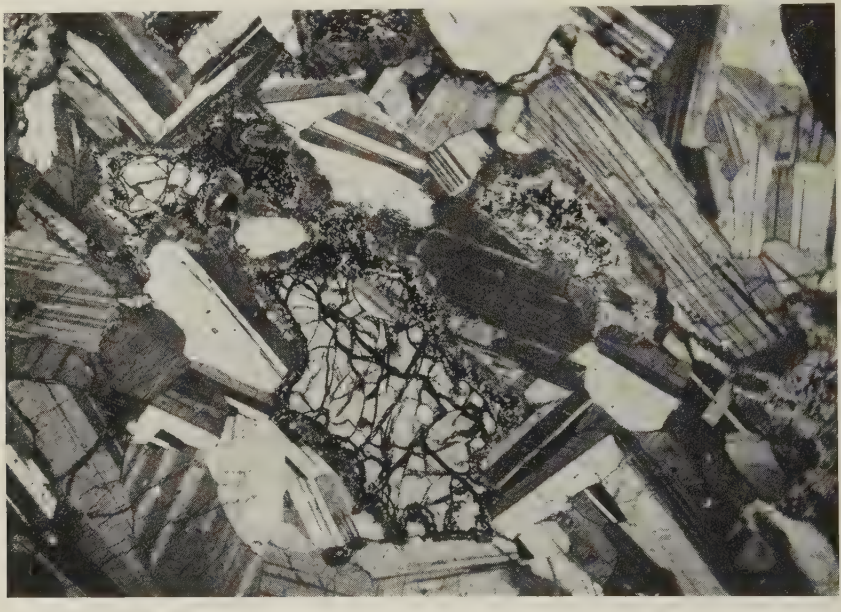

FIG. 2. Photomicrograph of thin section of anorthositic olivine-gabbro, to show kelyphitic coronas round olivine crystals. (Crossed nicols, ×24) West of Elm Street, St. Stephen.

In the olivine-norites (including the special varieties discussed below) the remarkable feature is the constancy in composition of the olivine. Determinations on material from all parts of the area show a divergence from $nZ=1.695$ (indicating a composition Fo_{66}) only within the limits of error of the immersion method ($\pm .003$, equivalent to a maximum error of 3 per cent Mg_2SiO_4). It will be noted, however, that the olivine in the norites differs in composition from that in the harzburgite. There is little divergence from An_{70} among the feldspars of the norites (excluding the quartz-gabbros, which contain more sodic feldspars) though here, since the refractive index curves are flatter, there is a greater possibility of error. The orthopyroxenes, on the other hand, exhibit wide variations, the range being from $nZ=1.675$ (enstatite, En_{90}) in troctolitic norite from the north-western quarry in Hanson Road, to $nZ=1.730$ (iron-rich hypersthene, En_{60}) in inclusions within the Rodgers pyrrhotite body. The magnesian types are colorless, but at a composition of En_{75} pleochroism becomes appreciable both in thin sections and powders.

A few examples of rocks in which augite is more abundant than orthopyroxene have been found; a mode illustrative of one such rock is given on p. 716. Such rocks, properly classed as olivine-hypersthene gabbros, appear to be subordinate to olivine-norites in bulk.

Troctolitic and anorthositic olivine-norites.—Banding in the norites takes the form of rhythmic alternations of concentrations of mafic constituents, particularly olivine, and felsic minerals. It ranges from barely perceptible streaking to the development of separate layers of melanocratic and leucocratic rocks. The extreme melanocratic type may be classified as troctolitic norite; a typical mode is given in Table 1. The rock is rich in rounded crystals of colorless olivine up to 2 mm. diameter, mantled with a small amount of enstatite and enclosed in a mosaic of coarse bytownite. Rocks rich in pyroxenes are rare and only one example was found.

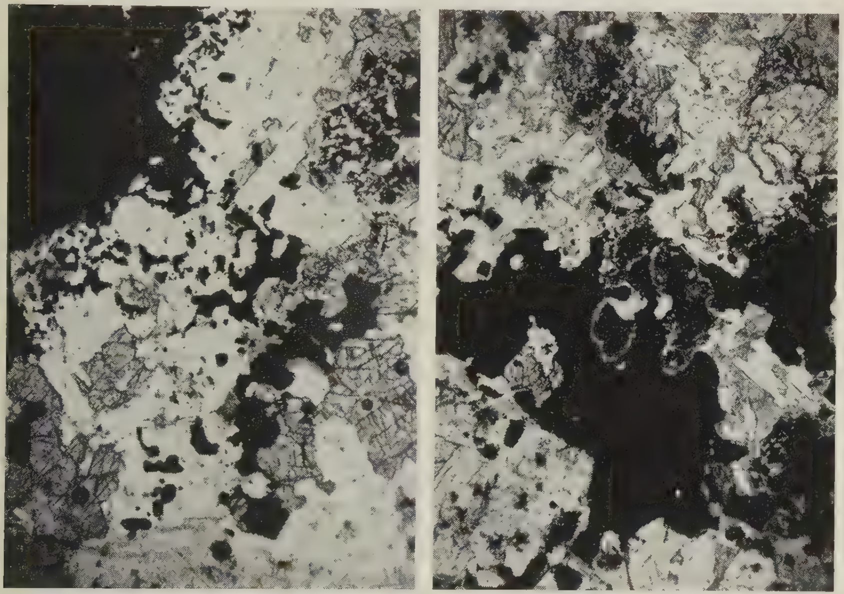

FIG. 3. Photomicrographs of thin sections of olivine-norite, to show replacement by magnetite and sulphides. In the left-hand photograph, the micropegmatite-like margin of a mass of pyrrhotite is shown ,and the preferential replacement of orthopyroxene by sulphides (Ordinary light ×30). In the right-hand photograph, partial replacement of rounded olivines by magnetite, surrounded by massive pyrrhotite is seen. (Ordinary light ×39). Rodgers prospect, St. Stephen.

The leucocratic rocks are largely composed of fresh labradorite-bytownite and these are classified as anorthositic norites. Such rocks not only enter into the banded portions of the complex, but also form substantial separate areas, for example between Elm Street and Cemetery Road. They are pale gray phanerides, mottled with evenly-scattered darker gray clots of ferromagnesian minerals, mainly olivine. None of the specimens studied show the achievement of true anorthositic composition, with over 90 per cent plagioclase; the nearest example carries 81.8 per cent. The olivine occurs as rounded crystals, enveloped by complex kelyphitic rims which are further described below. The plagioclase, which is coarsely twinned, shows neither zoning nor automorphism. The maximum crystal size is about 3 mm.

Relations with the invaded rocks; xenoliths.—The country rock into which the norites and harzburgite were intruded consists of quartz-chlorite-muscovite phyllites. Sliced specimens from near the mouth of Denny Stream, from Basswood Ridge, and also from Valley Road, remote from the norite, all indicate a similar metamorphic condition. The constituents include pale green, low-birefringent, optically negative chlorite, muscovite in flakes up to 0.05 mm. long, quartz, a little alkali feldspar, and evenly-scattered granules of magnetite, hematite and 'leucoxene.' The rocks are finely laminated; some bands being largely quartz-silt with a little chlorite; others being almost exclusively muscovite, well-oriented parallel to the bedding. All three main constituents may occur together in a band. The micaceous layers in some instances show small-scale drag-folding not exhibited by adjacent competent silt layers. The nearest exposure to the norite on Basswood Ridge failed to show any higher grade of metamorphism, and some of the xenoliths in the norite are in a similar condition.

Within the norite, however, on Basswood and Pomeroy Ridges, more intense metamorphism is exhibited by the large included masses of country rock. The micaceous minerals have been partially or wholly converted into deep brown biotite and the quartz has been recrystallized to a decidedly coarser grain-size. In the larger included mass on Basswood Ridge, abundant cordierite makes its appearance and certain bands carry hypersthene (En_{70}) associated with biotite showing $nZ = 1.640$, and a mosaic of quartz. The new ferromagnesian constituents in these free-silica hornfelses fail, in general, to maintain the preferred orientation of the chlorite and muscovite in the rocks from which they have been derived, but the lamination is preserved and evidently represents appreciable variations in composition. Small separate xenoliths of quartz-biotite-hornfels are found as far south as the cliffs of the St. Croix river between St. Stephen and Milltown. The further alteration of these hornfelses is considered in the next subsection.

Dark, fine-grained xenoliths from the St. Croix exposures prove to consist of pyroxene-granulite, composed of slender calcic labradorite crystals, with abundant drop-like granules of pale-green augite and magnetite. The close similarity of these rocks to the granulites in the gabbros of Skye (Harker, 1904, p. 115) and Ardnamurchan (Richey and Thomas, 1930, p. 230) suggests that they represent fragments of basic lavas incorporated in the norite. Such xenoliths are far less common than those of biotite-hornfels. A few examples have been found of diabase-like xenoliths having normal rather than granulitic pyroxenes, and carrying some olivine. It is not clear whether these represent a more advanced stage in the metamorphism of the granulites, or an early magmatic phase of the norites, broken up and incorporated in the main intrusion.

Doubt also attaches to the origin of angular white zenoliths up to 6 inches on the side, which occur in the Basswood Ridge road-cutting and elsewhere. The principal constituent is plagioclase of more sodic composition than in the adjacent norite (generally labradorite), forming granular aggregates of equidimensional crystals of 0.1 to 0.2 mm. diameter, engulfed in coarser plagioclase which also encloses crystals of dark green pleonaste up to 0.5 mm. long, and aggregates of white mica. A few crystals of amphibole and biotite are present in some examples, and clinozoisite occurs in dusky patches. Surrounding such inclusions there appears to be in all cases a zone, up to 5 mm. wide, exclusively composed of bytownite. It appears most probable that these rocks represent silica-poor hornfelses, but the necessary links with the country rock have not been found.

Marginal modifications of the norite.—Along the contact of the complex with the phyllites, and extending up to 100 yards from the contact, the norite gives place to quartz-hornblende-gabbro. Green and greenish-brown hornblende, both in coarsely crystalline and acicular varieties makes as much as half the rock. The composition of the amphibole is variable, as indicated by refractive index measurements ranging from $nZ' = 1.660$ in acicular types to 1.680 in massive varieties. Ferriferous biotite is also generally present, with nZ ranging from 1.635 to 1.650. The amphibole may contain cores of pyroxene and, occasionally, of olivine and the coarse type appears to have originated mainly from these minerals. The acicular type has also extensively replaced plagioclase. The feldspar is labradorite, appreciably more sodic than the feldspar of the norites. Quartz forms coarse interstitial areas. Rarely, a little micropegmatite is associated with it, and in one example, soda-orthoclase was found.

Similar rocks surround the large xenoliths or pendants of hornfels and here a stage representing the feldspathization of the hornfels, or the mingling of constituents derived from the hornfels with crystallizing feldspars can be recognized. The size of the biotite crystals as developed in the mixed rock is decidedly greater than in the hornfels and quartz is again the final product of crystallization. On the other hand the plagioclase—ranging in composition from An_{56} to An_{70}—is finer grained than in the norite, and of more obvious lath-shaped habit. Occasional larger crystals, comparable in size with those in the norite, appear as this rock is approached. Hornblende may be a major constituent. It is noticeable that both the marginal quartz-gabbros and those surrounding xenoliths contain apatite, a mineral rarely seen in the norites.

Adjacent to the pyroxene-granulite xenoliths the norite contains many inclusions, particularly of rounded augite granules. Coarse labradorite crystals along the margins extend from the norite into the xenoliths, and are of the nature of porphyroblasts in the latter. The norite itself is poor in olivine and unusually rich in hypersthene; it also carried more biotite than usual.

Norite-pegmatite; Aplites.—Small veins of leucocratic pegmatite are occasional features of the norites. A detailed study of one of these, from the road cutting on Basswood Ridge, showed that it consists of coarse plagioclase zoned outwards from about An_{50} to An_{32}. Green amphibole, also distinctly zoned with nZ' varying from 1.664 to 1.660, coarse biotite, partly altered to chlorite, quartz, calcite ($nO=1.657$) and apatite are also present. Ilmenite forms large crystals in the rock. An analysis of this rock is given in Table 2. Its interest lies in the fact that it indicates a trend towards the concentration of soda and of titanium in the residual liquors, but shows no concentration of potash.

Two aplite veinlets, respectively 1 and 3 inches wide, were investigated. Both proved to be microgranodiorites composed of zoned oligoclase, perthitic orthoclase, quartz, biotite and apatite. It seems more likely that these are associated with the granitic phase of the batholith than with the norites.

Hydrothermal alteration of the norite.—The descriptions given of the olivine-norite, and its troctolitic and anorthositic varieties, which together make up the greater part of the complex, will have made it clear that they are for the most part in a fresh condition, fresh olivine, for example, being the rule rather than the exception. Nevertheless, alterations yielding minerals containing the hydroxyl group, here described as hydrothermal, played a minor part in the evolution of these rocks. Alteration of the olivines takes several forms. The commonest is the development of a mineral showing perceptible pleochroism X, pale pinkish brown, Y, Z pale green; optically negative with low 2E, $nZ=1.589$, birefringence 0.03, one good cleavage, parallel extinction. This is tentatively assigned to bowlingite; the similarity of the mineral to talc in many of its properties (though not, of course, its color) may be noted. In this type of alteration, the olivine is generally pseudomorphed by the bowlingite together with a little magnetite. Fine-grained amphibole and antigorite also occur as pseudomorphs after olivine.

In the anorthositic norites, kelyphitic coronas are developed around fresh olivine crystals. These generally show an inner rim, next to the olivine, of low-birefringence, optically positive chlorite, probably pennine; and an outer rim of tremolite ($nZ' = 1.595$) adjacent to the plagioclase. No doubt the lime for the tremolite was derived from the feldspar, the magnesia from the olivine. Talc is developed in association with some of the coronas.

The orthopyroxenes are converted in places into amphibole for which the following data is representative: $nZ = 1.689$, birefringence 0.026, Z(pale brown) $> Y > X$(yellow), $Z \wedge c = 13°$, 2V near 90°. This data would be consistent with cummingtonite and it is quite likely that the lime-content is small. Other amphiboles, however, including those in the analysed olivine-norite, must carry appreciable amounts of lime in view of the difference between the normative and modal plagioclase. Fibrous amphiboles, though not common except in the quartz-gabbros, are occasionally found. In a few instances in the norites, actinolite with $nZ' = 1.633$ was noted. At least part, and probably a large part of the crystallization of the amphiboles took place after the consolidation of the norites. Joint faces in a quarry on Hanson Road yielded stellate clusters of green hornblende with crystals up to 3 inches long, showing $nZ' = 1.652$.

In the marginal quartz-gabbros it may be suggested that the formation of amphiboles and biotite was promoted by the incorporation of hydroxyl-bearing minerals from the country rock. It is unlikely, however, that the amphibole remote from the margins owes its origin to the same cause and there is similarly some biotite which must be assigned to the late stages of consolidation of the norite. This biotite is less ferriferous than that in the hornfelsed and marginal rocks, as indicated by its range of of refractive indices nZ between 1.608 and 1.612.

Sulphide Mineralization.—Pyrrhotite, in small amounts, occurs in all the rocks of the complex, including the hornfels xenoliths. Substantial concentrations are confined, as far as is known at present, to the localities listed on p. 717. The evidence is quite clear, as Dickson (1906) first demonstrated, that these concentrations were formed by replacement of the igneous rocks, textural features of which are preserved in the sulphides.

Polished sections of material from the Rodgers and Hall-Carroll prospects show, in doubly polarised light, that the pyrrhotite is a xenomorphic-granular aggregate of crystals ranging from 0.03 to 1.0 mm.

diameter. It encloses rounded magnetites (in some instances with plates of specularite) up to 0.5 mm. in diameter; these are cut by pyrrhotite veinlets. Intergrown with the pyrrhotite there are irregular groups of pentlandite crystals (soft, brass yellow, isotropic, showing cubic cleavage and triangular pits) and of chalcopyrite. Although elongate segregations of these minerals occasionally occur, the balance of evidence favours the view that they were deposited with the pyrrhotite, later than the magnetite.

Magnetite is only a minor constituent in the norites, occurring chiefly as a replacement of olivine associated with bowlingite or antigorite. In the mineralized ground, all stages may be found from incipient to complete replacement of olivine by magnetite, and there is no doubt that the rounded masses of magnetite in the massive sulphide bodies represent former olivines. When replacement of the olivine is effected by a mixture of antigorite or amphibole and magnetite, the latter mineral acquires the feathery texture to which Dickson has referred.

The sulphide mineralization was accompanied by the conversion of the silicates of the norite to hydroxyl-bearing minerals, chiefly actinolitic amphiboles, clinozoisite and chlorite. In the less intensely mineralized rocks, adjacent to the massive sulphide bodies, the pyrrhotite is surrounded by narrow zones of these minerals. There is some evidence of preferential attack by the sulphides on pyroxenes and amphiboles; certainly the olivines, whether altered to magnetite or not, were partially or wholly avoided. This type of alteration is illustrated by analysis 4888, Table 2; as compared with the norites, the increase in magnesia (in chlorite and amphibole) and of combined water will be noted. There is also a rise in the soda/lime ratio, representing the partial albitization, also evident under the microscope, of the feldspars. In this case, no rise in potash is found.

Within the sulphide lenses, the hydroxyl-bearing minerals have largely been eliminated by the advance of sulphide-replacement, and all that remains from the norite is a series of relics, mainly of anhydrous silicates. As at Sudbury, therefore, specimens can be obtained showing pyrrhotite corroding and replacing 'pyrogenic' silicates, and also of the sulphide accompanied by hydroxyl-bearing silicates. The rock analysed from the Rodgers prospect (Table 2, 5655), represents an approach to the massive sulphide stage. Combined water in this is no higher than in the normal norite; the feldspars are again albitized and here there is more potash than usual. The potash appears to be due to biotite which may represent the remains of hornfels xenoliths; no potash feldspar or sericite was found. Campbell and Knight (1907) give the following mineral sequence for St. Stephen: (1) Magnetite (2) Silicates (3) Sulphides. The present study suggests that (1) and (2) should be transposed.

Post-mineralization effects.—At the Rodgers prospect, a narrow north-south dike cuts through the sulphide body (*vide* Low, 1930, Fig. 3). This proves to be composed of elongate olive-green hornblendes which, with biotite, form half the rock, the remainder being made up of zoned andesine-labradorite (An_{45-55}), about 3 per cent quartz, calcite, apatite and a little pyrrhotite. This is classified as biotite-spessartite; no other example was found.

Shear-planes and joints cut the massive sulphides; on some of the former and on shear-planes in the norites, asbestiform minerals including anthophyllite and chrysotile are found. Data for an anthophyllite from Hanson Road quarry shows $nZ=1.646$, indicating, from the curve given by J. C. Rabbitt (1948, p. 295), about 13 per cent $FeO+Fe_2O_3+TiO_2+MnO$. Another anthophyllite, from the Hall-Carroll prospect, showed $nZ=1.629$. Brown-stained portions in both examples showed higher refractive indices. The crysotile has $nZ=1.556$ at the trial pit between the two branches of the C.P.R.; here, remains of olivines have also been replaced by this mineral.

PETROLOGY

The following petrological conclusions are drawn from the facts given:—

1. The constancy in composition of the olivine and bytownite suggest separation of these minerals from a magma of uniform composition. They did not, however, necessarily separate *in situ.*

2. The banding in the norites, parallel to the bedding and cleavage in the invaded rocks, cannot have been inherited from the sediments by metasomatism if argument (1) is accepted. In support it may be stated that where the interaction of the igneous complex and the sediments can be studied, mixed rocks rich in hydroxyl-bearing minerals are found; and that the large included hornfels masses showing bedding are surrounded by norites which fail to show banding. The banding requires further examination in the light of the work of J. W. Peoples (1936), L. R. Wager and W. A. Deer (1939) and F. Stewart (1947). In this area, where the banding is conformable to the bedding in the country rock, the hypothesis that the bands originated as horizontal layers prior to the regional folding might prove acceptable.

3. The magma was characterized by (i) low potash-content, a feature maintained throughout the rocks investigated; (ii) low water-content. Reactions around the margins of the intrusion involving chlorite and muscovite from the pelitic sediments augmented the available (OH) and led to the formation of hornblende- and biotite-bearing gabbros.

Feldspars formed in the mixed rocks were of more sodic composition than those in the norites, and were locally accompanied by a little orthoclase. Quartz from the pelitic sediments was dissolved and redeposited interstitially in the marginal gabbros. Apatite was similarly derived.

4. The low grade of metamorphism of the sediments adjacent to the complex may be ascribed either to a comparatively low magma- (or mush?)-temperature, or to the low content of volatiles in the magma, or both.

5. Incorporation of silica-rich hornfels may have contributed to the orthopyroxene-content of the norites, as Read (1935) demonstrated in the case of the Haddo House (Aberdeen) norite. The great variations in composition of orthopyroxene would be adequately explained on this basis. From this stage, crystallization *in situ* must be postulated for the main norites.

6. Mild hydrothermal alteration in the inner parts of the complex led to the development, by metasomatism in the solid, of cummingtonite from orthopyroxene; of tremolite from plagioclase; of bowlingite, antigorite, talc and pennine, with magnetite, from olivine. These changes probably involve addition of (OH) with only very local migration of other constituents.

7. Small pegmatites indicate a trend towards relative concentration of soda and titanium in residual fluids.

8. More intense hydrothermal alteration, perhaps due to residual fluids from lower parts of the complex, accompanied the very localized development, by metasomatism, of massive magnetite-pyrrhotite-pentlandite-chalcopyrite lenses. It is suggested that sulphidization eventually outstripped (OH)-metasomatism. The sulphide bodies are considered to be of 'pneumotectic' character (L.C. Graton and D. H. MacLaughlin, 1917).

9. A dike of biotite-spessartite, cutting through a pyrrhotite lens, represents the dying stage of basic magmatic activity here. Anthophyllite and chrysotile were formed on post-sulphide shear-planes.

10. Small aplites are believed, though not proved to belong to later granitic activity in the pluton.

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PARAGENESIS OF CUMMINGTONITE AND HORNBLENDE FROM MUURUVESI, FINLAND

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Abstract

Cummingtonite and hornblende occurring in homoaxial intergrowth in an amphibole-bytownite rock were separated and analyzed. Although the separation was not complete, the result is sufficient to show some essential features in the composition of both amphiboles as regards their different ratio [Mg]:[Mg+Fe] = mg; in the cummingtonite mg = 0.69, in the hornblende mg = 0.78. The variation interval of cummingtonite overlaps that of anthophyllite and, contrary to the statements of earlier investigators, the (Fe, Mg)-amphibole may be dimorphic. The nature of the polymorphism in this and related cases is discussed.

Every student of crystalline rocks probably has seen under the microscope a colorless amphibole in association with common green hornblende, occurring either as irregular spots or grown on the sides or at the ends of the hornblende prisms. The two amphiboles are always in homoaxial intergrowth, i.e. they have axes *b* and *c* and the face (010) in common. From analyses of the rocks and of the unhomogeneous amphiboles the colorless amphibole has been concluded to be cummingtonite, and its positive optical character is regarded an optical distinctive feature, though this is not always a reliable criterion. Sometimes anthophyllite occurs instead of or with cummingtonite. In other cases tremolite occurs in a similar manner, but cummingtonite is no doubt most common, being, in the amphibolite facies, the counterpart of hypersthene of the pyroxene-bearing mineral facies.

In his investigation of the gap in the solid solution series between anthophyllite-gedrite and cummingtonite on the one hand, and cummingtonite-grünerite and tremolite-actinolite on the other, Sundius (1933) gives analyses of anthophyllite and tremolite from Edwards, N.Y., in which these two minerals occur together. This, so far as known to me, is the only case where separation of lime-poor amphibole from lime-rich amphibole has been carried out. The separation of cummingtonite from hornblende never seems to have succeeded, although the assemblage of these two amphiboles is of wide occurrence in rocks. During my investigation of the Orijärvi area (Eskola 1914, p. 221), I tried to separate cummingtonite from the green hornblende present in cummingtonite amphibolite at the Orijärvi Mine, but this attempt resulted in failure due to their almost identical density.

Since 1914 I have been looking for more suitable material to separate cummingtonite and hornblende from one and the same rock, but invariably the very fine intergrowth did not encourage renewed attempts.

Finally, a rock was found that seemed worth attempting. It is a hornblende-bytownite rock from east of Jouhtenlampi, Murtolahti, parish of Muuruvesi, district of Kuopio, Finland, described by Wilkman (1938). The rock is millimeter-grained, aplite-like, consisting of tabular crystals of bytownite the interstices of which are filled with unhomogeneous, brown and colorless amphibole with some biotite and iron ore. It is said by Wilkman to form dikes in gneiss.

Wilkman had the rock analyzed by Dr. L. Lokka with the result given in Table 1, 1. Mr. E Helkavaara made another analysis of the rock (2). He also separated and analyzed the amphiboles under my supervision. The rock analysis was made in order to check Mr. Helkavaara's mineral analyses by comparing his rock analysis with another analysis of the same rock made by an expert analyst like Dr. Lokka. The agreement is satisfactory in view of the fact that the two analyses were not made on the same sample.

TABLE 1. AMPHIBOLE-BYTOWNITE ROCK

	1	2	Average	Mol. numb.	Norm			
SiO_2	51.18	51.35	51.27	8536	Q			0.72
TiO_2	0.58	0.47	0.53	66	Or			2.78
Al_2O_3	17.48	17.62	17.55	1722	Ab			15.31
Fe_2O_3	0.23	0.19	0.21	13	An			38.28
FeO	8.62	8.01	8.32	1158				57.09
MnO	0.09	0.01	0.05	7			Wo	2.91
MgO	9.12	9.64	9.38	2326	Di	5.67	En	1.78
CaO	9.03	9.22	9.13	1628			Fs	0.98
Na_2O	1.93	1.68	1.81	292	Hyp	33.89	En	22.16
K_2O	0.57	0.42	0.50	53			Fs	11.73
P_2O_5	0.03	—	0.03	2	Il			1.00
S	0.31	0.43	0.37	115	Mt			0.30
H_2O^+	0.79	0.76	0.78	433	Pyrrh			1.03
H_2O^-	0.08	0.08	0.08	—				
								41.89
	100.04	99.88	100.01					
d		2.908						98.98

The norm presents a close approximation to the mode of a norite of the same bulk composition as that of the analyzed rock in which amphiboles take the place of the pyroxenes and, besides some biotite, contain the (Fe, Mg)O in the rock. The composition of the normative plagioclase $Ab_{29\cdot 8}$. In the actual plagioclase were determined $\gamma=1.573$; $\alpha=1.564$, corresponding to the composition $Ab_{26}An_{74}$. The extinction angle in sections ⊥PM is 39°, corresponding to $Ab_{28}An_{72}$.

The intergrowth of the brown and colorless amphiboles is irregular, and the boundaries between them are gradual. Thus it is apparent from thin sections that the amphiboles are unhomogeneous. Decidedly more than a half of the amphibole is cummingtonite, but due to irregular distribution the exact proportions could not be determined. From estimates made I would say that the cummingtonite constitutes between 65 and 75 per cent of the amphiboles.

Table 2. Cummingtonite and Hornblende Analyses

	Cummingtonite			Hornblende		
	3	3*b*	3*b* Mol. numb.	4	4*b*	4*b* Mol. numb.
SiO_2	54.55	54.97	9153	52.41	51.59	8590
TiO_2	0.20	0.14	18	0.58	0.72	90
Al_2O_3	2.91	2.38	233	6.55	7.95	780
Fe_2O_3	0.19	0.06	4	1.14	1.51	95
FeO	16.37	17.25	2401	10.61	8.41	1170
MnO	0.02	0.02	3	0.01	0.01	1
MgO	21.57	22.11	5484	18.20	16.92	4196
CaO	2.77	1.84	328	8.67	10.95	1953
Na_2O	0.33	0.30	48	0.50	0.53	85
K_2O	0.13	0.10	11	0.37	0.47	50
H_2O+	0.80	0.80	444	0.85	0.87	483
H_2O-	0.03	0.03		0.04	0.04	
	99.87	100.00		99.93	99.97	
d	3.175			3.110		

For the separation of the two amphiboles, first the bytownite was removed by means of bromoform ($d=2.825$) in which it floated. Then the ore and other possible heavy minerals were removed with Clerici solution of $d=3.5$, then the solution was slowly diluted to $d=3.180$. At this density part of the colorless amphibole began to sink, and a considerable portion of it had sunk at $d=3.171$. The fraction $d=3.180$—3.171 was submitted to renewed separation at intervals 3.180—3.178, 3.176—3.174—3.172. The fractions between 3.178—3.176 and 3.176—3.174 were united and separated again. The main part sank between 3.177 and 3.174. This was taken as the cummingtonite fraction (analysis 3, Table 2).

At continued dilution of the solution a considerable portion of the hornblende was separated between the densities 3.119 and 3.095. The separation was repeated successively at $d=3.125$—3.120—3.118—3.114

—3.112—3.110—3.098—3.095—3.090. Microscopic examination showed that the fractions between 3.114 and 3.098 were richest in hornblende. These fractions were united and separated anew. The fraction 3.111—3.110 was collected and washed for analysis as the hornblende fraction (analysis 4).

The cummingtonite is colorless in thin section. The hornblende shows for γ greyish brown, β pale brown, and α yellowish, almost colorless. The angle $\gamma \wedge c$ in both is 20°–22°, and the dispersion $r < v$. The birefringence is just notably higher in cummingtonite, as appears from the interference colors in homoaxially intergrown crystals. The axial angle in both amphiboles is very large. The hornblende is decidedly negative, whereas 2V in the cummingtonite is so nearly 90° that it is hard to tell whether the mineral is negative or positive. Most determinations, however, indicated a positive character.

The separation was incomplete. According to a counting of the grains under the microscope the hornblende fraction would contain 21 per cent cummingtonite, but as the colored grains are more conspicuous than the colorless ones, the amount is probably greater, and was taken as 25 per cent. The cummingtonite was considerably purer, though grains of hornblende could be seen in it. Many of the colorless grains were speckled with pigment-like iron ore. The amount of hornblende in the cummingtonite portion was taken as 10 per cent. The analyses thus do not give exactly the compositions of these amphiboles. As, however, the materials contained only two amphiboles, it is possible to compute the composition of both kinds with closer approximation by stepwise eliminating the one and the other amphibole from the analytical figures, as described below.

From the original analysis of hornblende (Table 2, 4) 25 per cent cummingtonite (analysis 3) were subtracted. Thereafter 10 per cent of the residual figures, recalculated to 100 per cent and called analysis 4*a* (not quoted), were subtracted from analysis 3 and the residue was recalculated to 100 per cent (3*b*). This is the corrected analysis of the cummingtonite. Likewise 25 per cent of the percentage figures of 3*b* were subtracted from analysis 4. The residue, recalculated to 100 per cent, is quoted under 4*b* as the corrected analysis of the hornblende. The elimination could be continued in the same way, but as a trial showed that 3*c*, 4*c*, etc. would be practically identical with 3*b*, 4*b*, the calculation was not carried farther.

As may be seen from the molecular numbers, the figures under 3*b* and 4*b* agree fairly well with the Warren formulas for cummingtonite and hornblende, excepting the percentages of H_2O which are too low. Of course these figures can not be used for calculation of the amphibole formulas, as the factors 0.25 and 0.1 were based upon crude estimates only. Moreover, none of the two amphiboles was homogeneous, as appears from the facts that the intensity of the brown color in the hornblende varied, also in different parts of a grain, and that the density of the cummingtonite partly overlapped that of the hornblende.

It might be possible to achieve a better separation of the two amphiboles, e.g. by means of a magnetic separator and centrifuge. The analyses here published, however, give for the present purpose enough information as to the distribution of elements in conjugated lime-rich and lime-poor amphiboles. In the cummingtonite $[Mg]:[Mg+Fe]=mg=0.69$, in the hornblende $mg=0.78$. The hornblende, moreover, has taken up much more TiO_2, Al_2O_3, Fe_2O_3, Na_2O and K_2O in its structure, whereas the percentage of MnO is small in both. Furthermore the cummingtonite contains some Al_2O_3 and CaO.

Sundius (1933) claims that cummingtonite and anthophyllite represent different parts of the solid solution series between Fe- and Mg-amphiboles, and that a gap of a few per cent exists in the series from the ratio 60 per cent Mg-silicate upwards, the anthophyllite containing more Mg-silicate and the cummingtonite more Fe-silicate. The gedrites, however, may be richer in Fe-silicate, up to *c:a* 87 per cent, but instead they contain considerable amounts of alumina and ferric oxide. Sundius concludes that a content of Al_2O_3 and Fe_2O_3 of over 6 or 9 mol. per cent changes the monoclinic amphibole into the rhombic modification.

Rabbitt (1948) proposes, without stating the reason, that the name gedrite should be dropped and classifying the aluminous varieties with the anthophyllites, he extends the field of orthorhombic amphiboles up to ca. 75 mol. per cent Fe-silicate. On the Mg-silicate side he draws the boundary of the cummingtonite field at the same point as Sundius, though he remarks that the cummingtonite field may possibly be extended towards the Mg-corner on account of the cummingtonite from Sutherland, described by Collins (1942). It may be noted that the latter amphibole, in which $mg=0.68$, contains 5.02 per cent Al_2O_3 and therefore should not be given as much consideration as normal cummingtonites which are much poorer in alumina.

The cummingtonite from Muuruvesi is conclusive, showing $mg=0.69$ and less than 2.91 per cent Al_2O_3. This is the most important result of the present study. It is not the only example of such amphiboles known so far. In my Orijärvi work (Eskola 1914) I gave an analysis of an amphibole from cummingtonite amphibolite at the Julin Shaft, Orijärvi, with 30 per cent hornblende, showing $mg=0.68$, well above the solid solution limit according to Sundius.

We may state, therefore, that the fields of variation in the composition of cummingtonite and anthophyllite overlap at least within a range between the mg-values 69 and 60. This means that among the monoclinic and orthorhombic ferromagnesian amphiboles there exist examples exactly identical in composition. Therefore, Sundius' claim that cummingtonite and anthophyllite never present the relation of dimorphism, has to be modified. The same is the case with Rabbitt's conclusion that the anthophyllite and the cummingtonite series are not isodimorphous.

In some cases anthophyllite (or gedrite) and cummingtonite, when occurring together, are very different in composition. Such was the case in the garnet-gedrite-cummingtonite amphibolite from Isopää, Kalvola, Finland (Eskola 1936). In other cases, however, the monoclinic and rhombic amphiboles, in all probability, are identical in composition. This is especially the case when both are intergrown in the manner of laths twinned on {100}, as in many rocks from the Orijärvi area (Eskola 1914). A photomicrograph is in that memoir reproduced as plate V). No difference in refringence nor any other property can be detected except those dependent upon the different symmetry, the monoclinic laths showing oblique and the rhombic ones straight extinction. In this case, therefore, the dimorphism of the (Fe, Mg)-amphiboles would seem to be as close as that of the potash feldspar, with orthoclase and microcline, which seem to be identical in composition and all other respects except geometric symmetry. If Sundius' statement were right, cummingtonite and anthophyllite in such twin-like intergrowths would have different mg-values and different densities, so they could be separated one from the other. This will be worth an attempt!

I have (1936, p. 483) characterized the dimorphism in the (Fe, Mg)-amphiboles, supposing an identical composition for cummingtonite and anthophyllite, in the following words:

> "This kind of intergrowth, so extraordinarily similar to the twinning of the monoclinic amphibole, really would suggest a picture of their crystal structure somewhat like the following: Compare the silicon-oxygen bands of the amphibole with a group of soldiers standing in parallel rows. So long as all the rows are looking in the same direction they represent the monoclinic modification, but if every second row be commanded right about turn they will form an orthorhombic grouping. This would be equivalent to polysymmetry."

In the results of *x*-ray investigation of the structures concerned I cannot see anything that would materially contradict this rough model.

In most assemblages of anthophyllite and cummingtonite, however, these undoubtedly are different in composition. The differences are controlled by definite rules, the cummingtonite being especially richer in Ca and Fe and the anthophyllite in $(Al, Fe)^{3+}$ and Mg, while an identical composition is also possible. In other words, the diadochy (Strunz), or the ability of mutual replacement of the cations in both forms is different. The question is, whether the degree of diadochy of the cations is merely influenced by the coordination which obviously changes when two monoclinic unit cells turn over into one rhombic by doubling (Warren and Modell 1930). This apparently may be the case, as the lattice-energy thereby must change. Thus we may understand the relations in many such groups of crystal species in which two isomorphous series stand more or less strictly in an isodimorphic relation one to the other, and the vectorial properties (e.g. the optical ones) of the more symmetric modification can be derived from those of the less symmetric modification by means of the well-known Mallard equations, starting from the assumption that the one form represents submicroscopic polysynthetic twins of the other form. Further examples of such pairs of series are microcline and orthoclase, clinoenstatite and enstatite, clinozoisite and zoisite. The degree of isodimorphism is, of course, most perfect in such cases in which little or no isomorphic replacement of the cations takes place, as in the feldspars. Every mineralogist will at once see the different behavior concerning diadochy in the series mentioned.

Returning to the main subject of this paper it seems apparent from general petrographic experience that it is mostly cummingtonite and not anthophyllite that occurs in paragenesis with hornblende. Why? A natural explanation is that the monoclinic lattice of hornblende can only induce the crystallization of monoclinic (Fe, Mg)-amphibole. In its composition the latter may perhaps always be somewhat richer in iron than the hornblende, as we found in the present case, but it will vary according to the mg-ratio in the rock. It will be useful to have cummingtonite and hornblende separated and analyzed from many different rocks in which they occur together. At the same time it will be advisable to separate as many different fractions of both mineral species as possible. In this way only will it be possible to establish the succession of different members of the amphibole group, the variations of which no doubt will be very sensitive indicators of the conditions of origin and the laws of the metamorphic crystallization of the rocks.

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SYNTHETIC QUARTZITE*

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Abstract

Quartz sand (grain size 125–250 microns) immersed in weak aqueous Na_2CO_3 solution is converted to sheared quartzite (grain size 100–200 microns) after a few hours' exposure to temperatures in the range 230°–435° C., confining pressures between 5000 and 30,000 psi, and compressive loads between 32,000 and 103,000 psi. The principal feature of the experimental procedure is the use of a piston of rectangular cross-section and of a sand receptacle designed so that a biaxial strain is imposed on the quartz aggregate. The deformed prisms show shortening up to 57%, elongation up to 20%, and net decrease of volume up to 49%. Individual quartz grains have a maximum elongation of 2:1. Undulatory extinction bands are abundant, but no trace of deformation lamellae was found. There is little evidence of recrystallization. Both compaction and dimensional orientation are best developed in the upper range of temperatures and pressures used. Investigation of the lattice orientation (axis diagrams) of the deformed prisms shows random orientation for the most part. Calculation of the coefficient of correlation for each scatter diagram has also been carried out to test statistically this apparent randomness, but with inconclusive results.

Several similar experiments, using H_2O instead of Na_2CO_3, show a much lower degree of compaction and no trace of dimensional orientation; even, for example, after 7 days' exposure to 355° C., 30,000 psi confining pressure and 70,000 psi compressive load.

Introduction

Rock synthesis, for obvious reasons, has lagged far behind rock analysis as a field of geological investigation. Since successful synthesis depends in large part on knowledge obtained from analytical studies, this is not surprising and indeed is the logical order of events. As in analytical work, the approach may be either structural or compositional. Since the investigation reported here is concerned entirely with quartz, only structural considerations apply. The special theme is the behavior of quartz aggregates within a selected range of temperature and pressure. The following pages relate some details of the transition from uncemented sand aggregates to synthetic quartzite.

Background of the Problem

The facts regarding quartz orientation in tectonites are now well established as a result of work done by many petrologists from widely scattered parts of the world. I have recently summarized these investigations (Fairbairn, 1949) and attempted a synthesis of the conflicting hypotheses used to explain the data. Omitting details, it may be fairly stated that the factual side of the matter far outstrips the interpretative side. Experimental data are sorely needed if real progress is to be made. It was with this in mind that the present study was commenced.

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* It is a pleasure to acknowledge Professor Larsen's part in the development of the petrofabric background which has led to this paper. It was his scholarly interest in this field, then (1930) virtually unknown outside Austria and Germany, which first made me aware of its possibilities. Although not his own special interest, his encouragement at that time of this new work has in no small measure been responsible for my own continued interest in it.

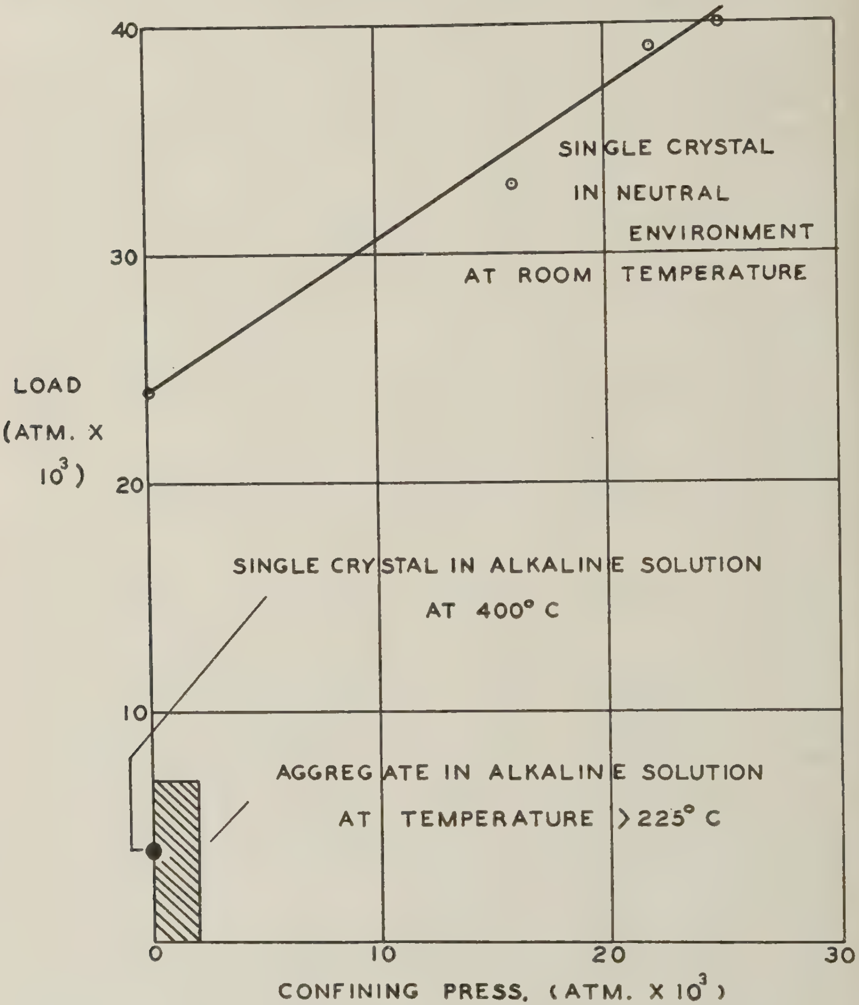

FIG. 1. Diagram showing strength properties of single-crystal quartz relative to experimental range used for cementation of quartz aggregates.

Enough work has been done on the deformation of single-crystal quartz to aid materially in the experimental procedure for deformation of quartz aggregates. Bridgman (1946) has made the most recent determination of the strength of quartz for confining pressures up to 25,000 atmospheres. Griggs and Bell (1938) showed that this high level of breaking strength could be drastically reduced by carrying out the experiment at elevated temperature (400° C.) in an alkaline solution environment. It was found that a crystal would fracture with one-sixth the compressive load required in Bridgman's experiment (Fig. 1). An additional contrast was the tendency toward crystallographic control of the shapes of the fragments in the Griggs-Bell experiment and lack of this feature in Bridgman's deformed material.

TABLE 1. EXPERIMENTS WITH QUARTZITE CYLINDERS

Expt. No.	Temp. (° C.)	Fluid Environment	Fluid Press. (psi $\times 10^3$)	Load (psi $\times 10^3$)	Time (hrs.)	Remarks
1	315	1% Na_2CO_3	15	13.5	1	Rupture after 1 hr.
2	315	1% $Na\ CO_3$	15	4.7	3	Rupture after 3 hrs.
3	315	1% Na_2CO_3	15	3.5	115	No rupture
4	315	2% Na_2CO_3	9	12.8	140	No rupture. Noticeable transfer of SiO_2 from upper to lower parts of cylinder.
5	150	10% Na_2CO_3	12	5.7	0	Ruptured immediately when load was applied.
6	Room	Air	0.0147±	30 (to produce rupture)	0	Value for load is average of 8 (Handbook Physical Constants—*G.S.A. Special Paper No.* **36,** 1942)

Table 1 summarizes the results of experiments with cylinders cut from quartzite[1] which confirm the contrast in the Bridgman and Griggs experiments. Experiment 6 is an average of 8 normal rupture tests. Experiments 1 to 5 show the effect of varying alkaline fluid concentrations, fluid pressures, and loads.

All of this work provides very definite evidence that the results of room temperature-dry deformation of quartz and quartzite give no indication of the actual confining pressures and compressive loads required to deform the mineral in the earth's crust.

Continuing this work, Griggs later constructed new apparatus and made a preliminary survey (Griggs, 1941) of the deformation of a number of minerals (single and in aggregate) in this temperature-pressure range. These investigations, discontinued because of the war, were taken up by me in 1946 through the generosity of Griggs in providing details of the equipment and a manuscript of his work as far as it had gone. Support has been provided in the succeeding years by a grant from the Geological Society of America, (Project Grant 466–55), to which society grateful acknowledgment is hereby made.[2]

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[1] Massive, fine-grained material, locality unknown. All cylinders have the same orientation.

EXPERIMENTAL PROCEDURE

Full details of the equipment used are in process of publication elsewhere. It is necessary to include here only a description of the receptacle in which the deformation of the quartz sand is accomplished. This consists of a slot with vertical, parallel sides cut in one end of a piece of drill-rod steel (Figs. 2, 3) into which fits a Carboloy piston of similar shape. The piston does not occupy the full length of the slot, the ends being blocked with shaped copper pieces as shown. The space between the copper end pieces is filled almost to the top with the quartz aggregate. A close-fitting copper cylinder is inserted around the steel piece, project-

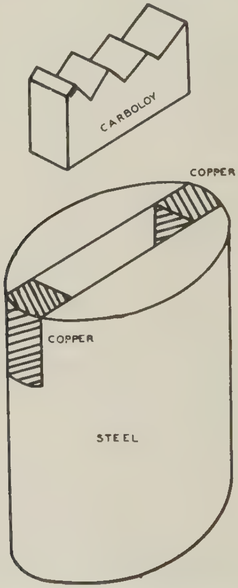

FIG. 2. Receptacle used for quartz aggregate experiments. See also Fig. 3 (*a*) and (*b*).

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[2] Invaluable mechanical assistance by John Solo of the Department of Geology Shop solved many problems which plagued the investigations. Laboratory assistance at various times by J. B. Thompson, Jr., M. C. Wittels, D. L. Kendall, and R. H. Stebbins is also acknowledged.

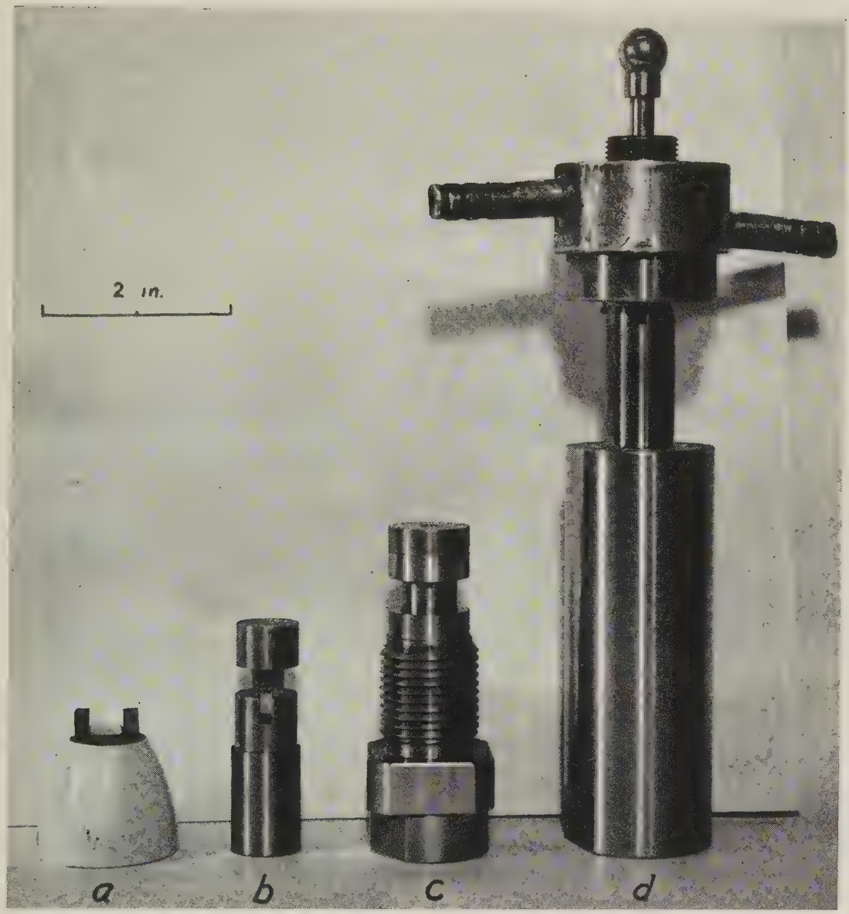

FIG. 3. Pressure chamber and receptacle assembly. (*a*) Copper end pieces enclosing a prism of synthetic quartzite. (*b*) Copper sheath at base; steel receptacle, center; carboloy piston, top. (*c*) Assembly for sealing base of pressure chamber. (*d*) Pressure chamber showing outer piston (top) and water-cooling jacket.

ing above the top. The space above the slot is filled with Na_2CO_3 solution. The piston is then inserted in the top of the slot, excess solution being expelled at the contact with the copper cylinder. By this means air is displaced from the receptacle. The cylinder serves further (1) to hold in place the copper pieces in the slot, and (2) to protect the receptacle from convection currents in the main body of the bomb. The loaded receptacle is now placed in the bomb and the experiment may proceed.

Temperature is first raised to the desired level and the fluid pressure adjusted to the required setting. Finally a load is applied through the piston assembly shown in Fig. 3 (*d*). This may be done in one stage, or by increments at set time intervals. Most of the experiments were run until the dial gage indicated that shortening of the specimen had ceased. Many were run beyond this time.

The grain size of the quartz aggregate was held between 125 and 250 microns so that optical study of the deformed material would not be unnecessarily difficult. No attempt was made to select rounded grains. The quartz sand was obtained by disaggregation and sizing of selected specimens of the Nepean sandstone from Ottawa, Canada (Cambrian). The grains are mostly angular and include both equant and inequant shapes in approximately equal amounts (Fig. 4).

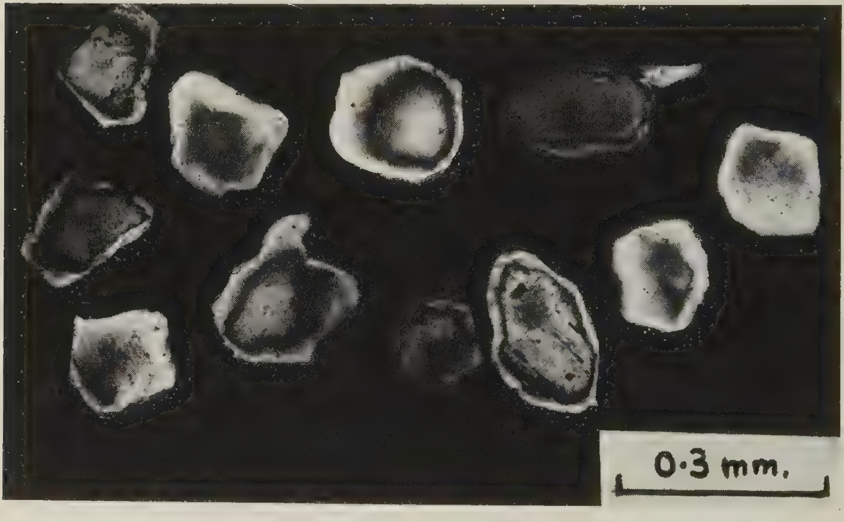

FIG. 4. Photomicrograph of typical disaggregated grains of the Nepean sandstone used in the experiments.

Upon completion of the experiment the receptacle is taken from the bomb and the Carboloy piston and cylindrical copper sheath are removed. The walls of the slot are then carefully sawed off at the base, thus exposing the deformed specimen, which is easily lifted out. Its height and length are then recorded and compared with the original dimensions. Width remains constant because of the design of the receptacle (Figs. 2, 3). A thin section is prepared of the side of the specimen parallel with the vertical wall of the slot. This slide is then examined for visible fabric details and the orientation of the quartz axes is determined with a universal stage.

The equipment used gave temperatures up to about 450°C. in the space occupied by the receptacle. As the temperature in the receptacle is not

TABLE 2. EXPERIMENTS USING 5% Na_2CO_3 SOLUTION (pH = 8)

Expt. No.	Temp. ° C.	Fluid press. psi×10^3	Load† psi×10^3	Time of Expt. (hrs.)	Est. time to produce 75% of shortening	% Shortening	% Elongation	Cementation	Dimensional orientation	% Volume decrease	Statistical classification of axes orientation		Remarks
											Random	Significant	
1	150	30	70	19	4.5	43	nd	None	None				Aggregate finely granulated
2	190	30	70	63	40	36	nd	None	None				Temperature threshold for cementation is at 200° C. approximately
3	230	0	70	15	3	11	nd	None	None		X		No fluid pressure used in excess of that generated at the temperature of the experiment
4	230	5	70	9		nd	nd	Moderate	None		X		Time-shortening* curve erratic
5	230	15	70	22	18	36	nd	Moderate	None			X	Temperature threshold for dimensional orientation is at 230° C. approximately
6	230	27	70	66	46	41	nd	Excellent	Moderate		X		
7	230	30	70	9		43	10	Moderate	Moderate	37	X		Time-shortening* curve erratic
8	300	25	14	23 days									Quartz grains transferred to non-stressed parts of receptacle
9	315	7	32	10	10	nd	nd	Excellent	Moderate		X		Shortening continues for longer period than in Expt. 11 because of relatively low fluid pressure
10	315	13	32	19	11	nd	nd	Excellent	Moderate				Thin section missing
11	315	26	32	71	5	nd	nd	Excellent	Moderate			X	Prolonged experiment (71 hrs.) does not visibly improve the dimensional orientation
12	315	30	103	24	6	nd	nd	Excellent	Moderate			X	Load applied by stages. Specimen indistinguishable from preceding experiments at same temperature
13	355	5	70	27	3	36	nd	Slight	Slight		X		Short time (3 hrs.) for 75% of shortening related to poor cementation, resulting from low fluid press.
14	355	15	70	24	9	27	nd	Excellent	Moderate		X		
15	355	25	70	64	22	nd	nd	Excellent	Excellent			X	
16	355	28	70	35		45	nd	Excellent	Excellent			X	Time-shortening* curve erratic
17	435	10	70	9	10	nd	nd	Excellent	Moderate		X		
18	435	20	70	8.5	6	57	20	Excellent	Moderate	48		X	
19	435	20	70	31	5	48	13	Excellent	Moderate	41		X	Load applied gradually. Specimen indistinguishable from Expt. 18, even though run was prolonged almost fourfold
20	435	10	103	9	8	56	15	Excellent	Slight	44	X		Temp. control poor
21	435	20	103	24	8	53	15	Excellent	Excellent	43	X		Expts. 20–22 (103,000 psi load) tend to develop more fracturing of the specimens than in those under smaller load
22	435	20	103	11		55	20	Excellent	Moderate	49		X	No improvement of dimensional orientation is seen
23	435	25	70	3.25	.4	56	15	Excellent	Excellent	49			Saturated solution used (pH = 11)

* This is probably due to non-uniform friction characteristics of the outer piston.

† Total load applied at beginning of experiment, except where otherwise noted under "Remarks."

measured directly an error of perhaps ±5° is unavoidable. Therefore control of the temperature is unwarranted to limits closer than ±5° and was not attempted.

Fluid pressures between 5,000 and 30,000 psi were maintained. The sensitivity of the Bourdon gages used is ±1%. Automatic control was not necessary as variations of several thousand psi had no observable effect on the course of the experiments.

Compression loads were used which by calculation should transmit between 35,000 and 115,000 psi to the receptacle. The fraction actually transmitted probably does not exceed 90% of these values on the average. The limits might thus be set at 30,000–100,000 psi.

For most of the experiments aqueous Na_2CO_3 solutions of various concentrations were used. Na_2SiO_3 was used successfully in some runs. H_2O was also tried in a number of experiments.

The time for individual experiments varied between 117 days and $2\frac{1}{4}$ hours. For the most part the runs were of 24 hours' duration or less.

RESULTS OF EXPERIMENTS

Sodium carbonate is one of a number of non-corrosive salts which in aqueous solution provide the alkalinity necessary for these experiments. A 5% solution ($pH=8$) was used for most of the runs, since very weak solutions prolong the experiments unnecessarily. On the other hand, under the conditions of the experiments, saturated solutions (Expt. 23, Table 2) react somewhat too rapidly with the quartz for accurate appraisal of the results.[3] Table 2 therefore is largely concerned with data using a 5% solution. The experiments are arranged in order of increasing temperature; within each temperature group the order is that of increasing fluid pressure.

Application of the load in increments (Experiments 12, 19, 22) rather than in one stage makes no observable difference in the deformed specimen, other than increasing the time required for compaction.

A minimum temperature of about 200°C. is necessary for appreciable cementation of the aggregate. In combination with this an appreciable excess fluid pressure[4] is needed. (Experiments 1–4), as well as a compressive load. Cementation was not obtained with less than about 5,000 psi fluid pressure. The smallest compressive load used (32,000 psi, Expts.

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[3] This results in transfer of the grains from the stressed space beneath the piston and redeposition on the walls of the receptacle. Experiment 8 (Table 2) shows that, even with a 5% solution, a prolonged run results in loss of the material. The rounded edges of the deformed specimen shown in Fig. 3(a) are evidence of the beginings of this transfer.

[4] That is, in excess of the relatively small pressure generated at the given temperature level.

9–11) does not represent the minimum possible load. An investigation of the minimum compressive load was not feasible, as the error arising from frictional forces in the external piston sleeve would become disproportionately large. It was not possible either to test effectively the factor of increasing time, since there there is continuous solution of the quartz grains under the piston and re-deposition in unstressed areas of the receptable (e.g. Experiment 8). The time required for 75% of the compaction can be estimated[5] and is a somewhat better basis for comparison of experiments than the actual time of a run (compare time columns in the table).

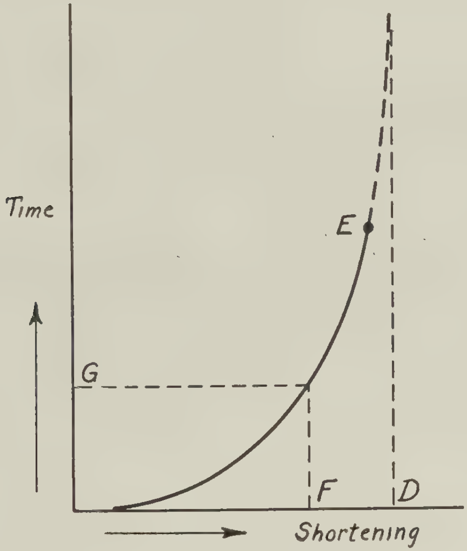

FIG. 5. Diagram showing method of approximating the time required for 75% of shortening to take place. (See footnote in text)

Dimensional orientation of the deformed aggregates first becomes noticeable at about 230°C. There is elongation of the grains normal to the load and parallel with the length of the slot in the receptacle (Fig. 6). Maximum ratio of the long and short axes is about 2:1. Although some grains in the undeformed sample were inequant to almost this extent, it is certain that rotation of these into parallelism is an inadequate explanation of the observed dimensional orientation. There is a net decrease

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[5] The time vs. shortening curve for any experiment is approximately exponential and the time required for essential completion of shortening can be estimated even where an experiment is not run to completion. For example, in Fig. 5, *E*—experiment stopped here; *D*—estimated total time for experiment if completed; *F*—75% of shortening represented by *D; G*—time required for 75% of estimated total shortening.

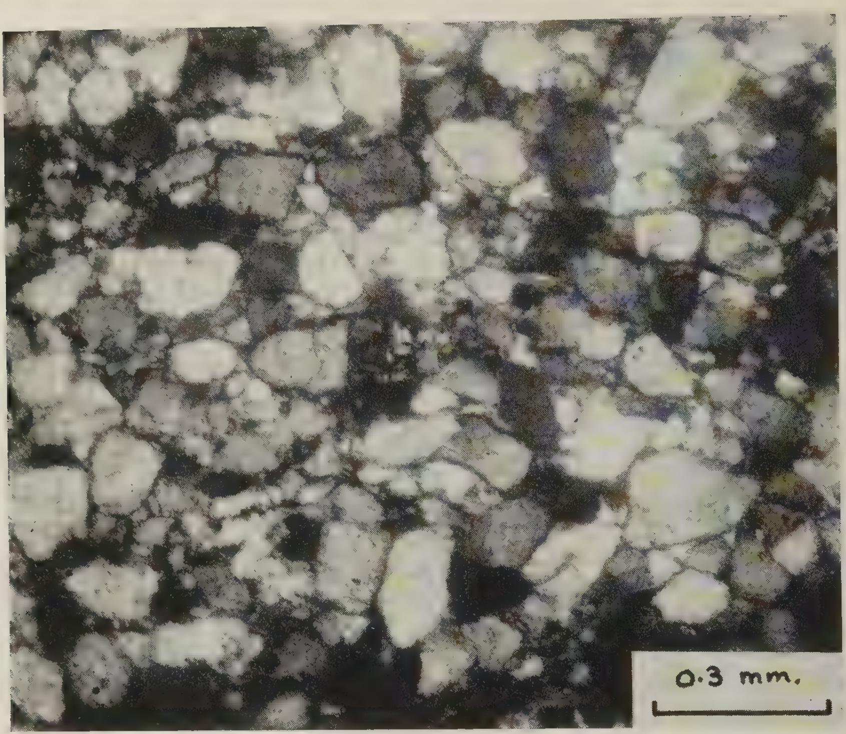

FIG. 6. Photomicrograph of synthetic quartzite showing elongation trend parallel to long axis of slot in the receptacle (east-west). Note non-uniform grain size and brecciation.

in range of grain size from 125–250 (undeformed) to 100–200 microns (deformed) and considerable less angularity in individual grains after deformation than before (Fig. 7).

Examination of thin sections shows the consolidated, deformed material to be a sheared quartzite (Fig. 6). There is considerable brecciation around some of the grains and all show undulatory extinction. No trace of deformation lamellae has been found. Axes diagrams have been made by means of the universal stage for all deformed aggregates from which a thin section could be made. No definite orientation pattern could be determined by inspection alone, although, as Table 2 shows, a wide range of variables was tried out. To test statistically the degree of randomness the coefficient of correlation was computed for each point diagram, as outlined by Chayes (1949). This study indicates that about one-half the diagrams are probably random (isotropic), the other half probably significant (anisotropic). As this statistical classification does not, however, show any rational relation with the corresponding experi-

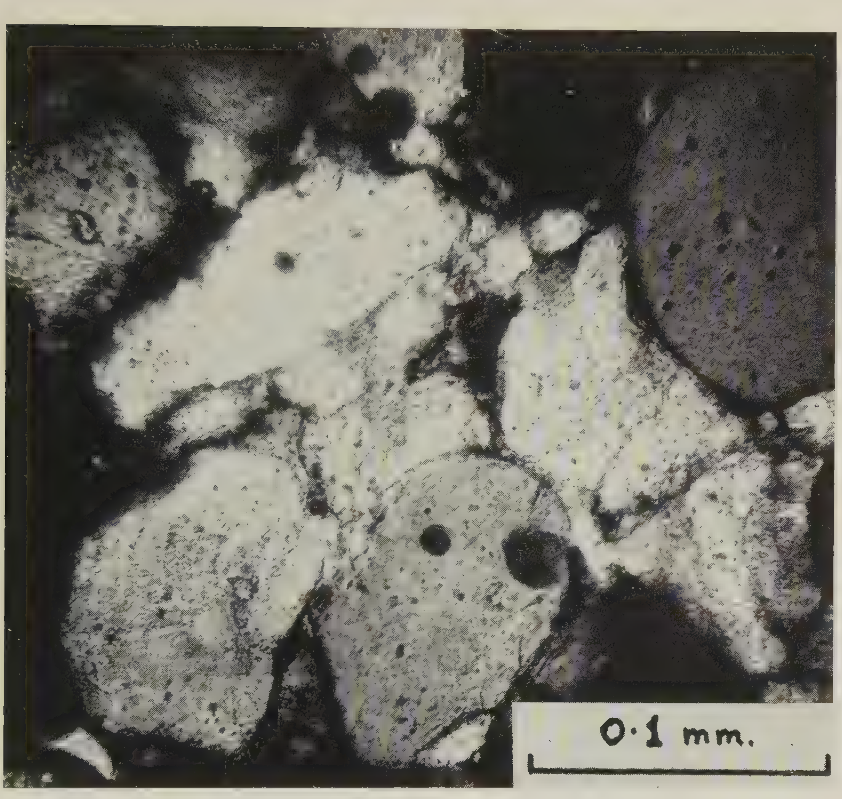

Fig. 7. Photomicrograph of synthetic quartzite at high magnification showing rounded grain outlines and interstices filled with small grains.

mental conditions (see Table 2) the analysis is indecisive as matters stand at present. It is possible, however, that the lattice orientation, as determined with the universal stage, does not represent the entire deformation story. Many grains in these sheared quartzites escape measurement because of their minute size. Until their part in the stress-strain picture is known the final word regarding the overall character of the axes orientation can not be said. It is hoped that this additional study can be successfully carried out using an improved type x-ray spectrometer.

With respect to the optically measurable, larger grains in the deformed quartz aggregates it would appear that changes in the fabric involve transfer of material from points of high stress to points of lower stress, resulting in elongate grains which tend to be arranged perpendicular to the direction of shortening. No important grain rotation or change in the axes orientation seems to take place. The process is essentially Riecke's principle applied to a polycrystalline aggregate. The conditions

TABLE 3. EXPERIMENTS USING H_2O (1–3) AND Na_2SiO_3 (4–5)

Expt. No.	Temp. ° C.	Fluid press psi×10^3	Load† psi×10^3	Time of Expt.	Est. time to produce 75% of shortening	% Shortening	% Elongation	Cementation	Dimensional orientation	% Volume decrease	Remarks
1	300	20	14	117 da.				None	None		Time-shortening* curve erratic. See footnote to Table 2
2	355	28	70	3½ da.	20 hr.	36	nd	Slight	None		
3	355	30	70	7 da.	22 hr.	36	nd	Slight	None		
4	275	20	70	4 hr.	1 hr.	45	12	Excellent	None	40	Load added by stages
5	435	25	70	2¼ hr.	4 min.	51	23	Excellent	None	40	

* This is probably due to non-uniform friction characteristics of the outer piston.

† Total load applied at beginning of experiment except where otherwise noted under "Remarks."

of the experiments have not produced, therefore, any of the lattice orientation patterns commonly associated with quartz tectonites.

A few experiments with water glass (Na_2SiO_3) indicate behavior like that observed with saturated Na_2CO_3 (Table 3, Table 2, Expt. 23). No extended series of experiments was carried out, however, as it is believed to be $(OH)^-$ concentration, not the compound itself, which accounts for the effects observed.

A number of experiments were run using H_2O as a fluid environment. No satisfactory cementation was obtained by this means (Table 3) and no dimensional orientation was observed. Only with difficulty could thin sections of this deformed material be made. Possibly the slight cementation observed in two of the experiments arises from traces of OH^- in the bomb as a result of inadequate cleaning. It is concluded that absence of $(OH)^-$ is the key to the matter.

Discussion of Results

It has already been stated that the failure to produce evidence of lattice orientation in these synthetic quartz fabrics may be total, or only partial, depending on what additional evidence may be obtainable by x-ray methods. At any rate the large, optically measurable grains lack evidence of lattice orientation. The development of the parallel dimensional orientation goes on during the stage of cementation and elimination of voids. After this stage is completed, the fabric of the synthetic quartzite is frozen into a permanent pattern which remains unchanged as far as the present group of experiments is concerned. To test this from another angle, thin, parallel-sided plates of natural quartzite[6] were subjected to the same range and combination of variables as listed in Tables 2 and 3. In the half-dozen experiments run there was found to be, without exception, no observable change in lattice or dimensional orientation of the plates, thus confirming the conclusions reached for the synthetic material. The conditions of the experiments are therefore inadequate for reproduction of the girdle orientation patterns of tectonite quartz.

Higher temperature might produce lattice orientation, but as the present equipment is not designed for much higher temperatures and pressures than those already used, this aspect of the investigation is problematical. Lengthening the time of an experiment carries with it two serious restrictions, (1) the short active life of the external piston, and (2) the continuous solubility of quartz in alkaline carbonates (enhanced by the pressure differential of the experiments).

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[6] The Lorrain quartzite (Huronian), Sudbury District, Ontario, was used for this purpose.

Natural quartzites are known which have random lattice orientation (Phillips, 1937), but without more detailed information as to their tectonic history it would be unwise at present to attempt correlation with the lattice orientation of the synthetic product described in this paper. Fairbairn (1949, pp. 167–169) has outlined some hypothetical stages of recrystallization which, if applied here, would involve paratectonic recrystallization and an intermediate state of dimensional orientation, with no change in lattice orientation. The deformed material, however, is poorly recrystallized, indicating possibly that granulation has outstripped the recrystallization process.

The conditions of the experiments impose a nonrotational strain, or restricted type of transport on the material (Fairbairn, 1949 Chap. 17). As already seen, this does not reproduce the common girdle orientation of tectonite quartz. If rotational strain is necessary for production of these girdles, the only feasible way to impose it on the experimental material is to apply a torque in addition to the normal compressive load. Some work has been commenced in this direction, but since *x*-ray analysis of the deformed material appears to be necessary, no thorough study has yet been made.

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THE AQUEOUS EMANATION FROM PARÍCUTIN VOLCANO[1]

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Abstract

An estimate of the quantity of the water emitted by the crater vent of Parícutin volcano gives 17,000 tons per day, compared to an average daily emission of lava of 100,000 tons from the lava vents. This quantity of water is believed to be larger than the amount of water one could reasonably expect from the magma rising in the eruptive conduit, and suggests a considerable dilution of magma emission by vapors derived from meteoric waters. The apparent differences between crater vent and lava vent emissions bear out this idea. The imposing eruptive column of a volcano would, in such a case, be no measure of the quantity of magmatic water evolved by a crystallizing magma. It is suggested that groundwaters activated by an intrusion into the zone of these waters may play an important role in the hydration and the alteration of wall rock.

One of the problems facing the student of ore deposits is to find a source of water sufficient to account for the huge masses of hydrated and altered rock frequently accompanying certain types of ore deposits. As an example, the propylitization of andesite wall rock accompanying the veins of such a deposit as the Comstock Lode requires the fixation of about 50,000,000 tons of water per cubic kilometer of hydrated rock.[2] In addition to water so fixed, an additional large quantity must be postulated to account for the leaching that evidently occurred during propylitization. Since propylitized or similarly altered rock frequently occupies a number of cubic kilometers, the quantity of water necessary to achieve this alteration must be very large.

Some geologists see an ample source of water for this and similar purposes in the emanations from magmatic bodies and cite the evidently great quantities of steam from a volcanic apparatus as a measure of this tremendous supply. (Kemp: 1906, 1908; Phemister: 1934). Some estimate or measure of the quantity of water vapor emitted, therefore, is desirable, as well as some indication of its probable source.

The eruptive column is, indeed, one of the most impressive features of an active volcano. It consists principally of water vapor, mixed with other gaseous emissions, and is frequently heavily charged with ash. Even when no visible vapor column is apparent because of its superheated condition, the emission of water vapor manifests itself, if atmospheric conditions are propitious, by the development of huge cumulus clouds above the volcano.

Under favorable circumstances an estimate, based on observable factors, of the quantity of water vapor evolved by a volcanic apparatus,

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[1] Published with the permission of the Secretary of the Smithsonian Institution.

[2] Based on analyses of andesite and propylite from the Comstock Lode. *U. S. Geol. Exp. 40th Parallel*, **III**, 90 (1870).

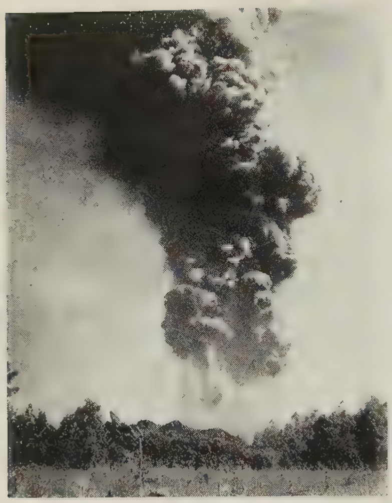

FIG. 1. Heavy eruptive column characteristic of the early Quitzocho period. Parícutin volcano. March 24, 1943.

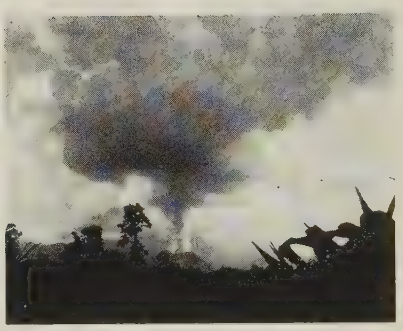

FIG. 2. Condensation of water vapor to heavy cloud. Zapicho vent, Parícutin volcano. Dec. 6, 1943.

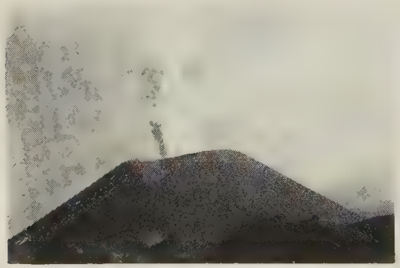

FIG. 3. Eruptive column of steam at time of observations. Parícutin volcano. May 28, 1945.

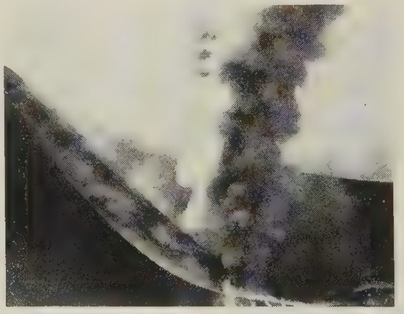

FIG. 4. Crater vent at time of observations. Parícutin volcano. May 27, 1945.

can sometimes be made. The new Mexican volcano Parícutin has offered such an opportunity.

VOLCANIC PHASES AT PARÍCUTIN

The activity of Parícutin volcano may be divided into three phases. The first, or Quitzocho, period beginning with the original outbreak of the volcano on February 20, 1943, and continuing to October 19 of the same year, involved the building of the volcanic cone. The eruptive activity was characterized by very heavy vapor emission, yielding a majestic eruptive column (Fig. 1), heavy ash fall but with relatively small and erratic lava flows. The second, or Zapicho, period began with the outbreak at the Zapicho vent and the building of the adventitious cone, Zapicho, at the northeast base of the main cone. Zapicho yielded a large lava flow, but very little ash, and invisible eruptive vapors. The vapor emission from Zapicho, however, sometimes manifested itself as huge condensation clouds high above the vent (Fig. 2). During this phase which continued from October 19 to January 8, 1944, the main crater emitted lazy cumulus-like clouds of vapor. The third, or Taqui, period beginning with the outbreak of the Taqui lava vent at the southwest base of the cone on January 8, 1944 was characterized by almost constant but variable activity from the crater (Fig. 3) vent, but with little ash and almost constant emission of lava from one of three lava vents at the base of the cone.

TABLE OF PARÍCUTIN ACTIVITY 1943–1946

Stage	Duration	Ejectamenta Solid Av. per day	Activity
Quitzocho	Feb. 21–Oct. 19, 1943	315,000 tons	Heavy ash laden eruptive column; moderate lava flows.
Zapicho	Oct. 19, 1943–Jan. 8, 1944	350,000 tons	Large lava flow; moderate but constant vapor emission from adventitious vent.
Taqui	Jan. 8, 1944–46	100,000 tons	Large lava flows; slight ash fall; variable vapor emission.

Estimates have been made of the solid ejectamenta, based on the extent of lava and ash cover measured on air photographs and estimates of thickness of lava and ash in the field. The weight per unit volume of ash and lapilli was determined by weighing measured quantities of these products and found to be 1500 kgs/m^3 and 1250 kgs/m^3, respectively. The porosity of the lavas, as estimated from sections exposed in crevasses in the flows, was estimated at 50 per cent. The following table shows the characteristics of the three eruptive phases of Parícutin and the average daily quantity of ejectamenta.

During the Taqui phase it was feasible, at times, to ascend the cone. On one such ascent, on May 27, 1945, at which time the crater was in average eruption for this period, neither unusually violent nor greatly reduced, (Fig. 3) conditions in the crater offered requisites for a satisfactory estimate, based upon directly observable factors, of the quantity of water vapor being emitted by the crater vent. A continuous column of white steam arose from a plainly visible vent near the north edge of the crater with a loud grating roar (Fig. 4). This chimney-like orifice had an estimated diameter of 2 meters. Although the steam emitted was, in general, condensed, there were brief intervals during which it issued as superheated, invisible vapor, but with no evident abatement in force or volume as manifested by the uninterrupted roar accompanying its escape. This condition suggests that the temperature of the escaping vapors had an average temperature of about 100°C. A few scattered bombs were carried out by the rushing vapor column, the highest of which took 10 seconds to fall back to the level of the vent, from which may be calculated, from the relationship of gravity acceleration, a "muzzle velocity" of the bombs, and therefrom the velocity of escape of the vapors of 100 meters per second. These data are, perhaps, as good as can be obtained considering the insurmountable difficulties in making precise measurements. From the density of steam at 100°C. the quantity of water vapor emitted was calculated as 12 tons per minute, or 17,000 tons per day.

In addition to the water escaping from the crater vent, there were additional quantities emitted from the lava vents, the lava flows, and from the slopes of the cone. No satisfactory basis for calculating these amounts is offered, but the quantity apparently was considerably less than that issuing from the crater vent.

With the daily emission of lava during the Taqui phase of about 100,000 tons, the apparent water content of the lava, assuming this emission totally derived from the lava concomittantly extruded, exceeded 17 per cent. This is consideraly more water than is generally believed to be the probable aqueous content in magma. The solubility of water in granite magma has been found to range up to as much as 10 per cent (Goranson, 1931). It is generally supposed that basaltic magma is less hydrous than granitic. At 1200° C., which is the temperature of Parícutin lava as extruded (Zeis, 1946), and 980 bars, corresponding to 3½ kilometers in depth, the maximum solubility of water in granitic magma is 4.8 per cent (Goranson, *loc,. cit.*). Gilluly (1937) concludes that basaltic magma contains upward of 4 per cent, and Phemister (1934) 5 per cent of water. Friedlaender (1926) calculated the emission of water vapor for a 20 hour period beginning on April 8, 1906, at Vesuvius at 1.75 km^3, compared to 0.20 km^3 of primitive solid ejectamenta in the form of lava and ash. Friedlaender's estimate of a crater throat diameter of 500 meters is undoubtedly many times too large and the estimated quantity of aqueous emission correspondingly too high. His figures, however, do indicate that the aqueous emission, compared to the solid ejectamenta, is inordinately high for a magmatic source. Bowen (1928), however, argues for a comparatively anhydrous character of basaltic magma.

The gas discharge from volcanos is frequently cited as evidence of a high water content of magma (Gilluly, *loc. cit.*, Phemister, *loc. cit.*). Kemp (1908) goes so far as to state "Volcanos and all other outbreaks visible to us display vast quantities of steam emitted in such a way as to leave no doubt of its derivation from the molten magma." The high ratio of emitted water to extruded lava arrived at above, however, suggests that, in the case of an imposing eruptive column, the magmatic emanations have been diluted by groundwaters. A further indication of this dilution is the strongly contrasting character of the emanations from these lava and lava vents and of the emissions from the crater vents. Samples of emanations from the lava vents have been collected but have not yet been quantitatively analyzed. Preliminary examinations indicate a high content of hydrochloric acid, with lesser amounts of sulfur dioxide, sulfur trioxide, hydrogen sulfide, carbon dioxide and other minor constituents. The presence of flames in the "hornitos" above the flowing lava indicates the presence of combustible gases, perhaps hydrogen. These emanations are so noxious that it is impossible for one to approach an active lava vent from the leeward side. On the other hand, the emissions from the crater vent, unless the activity is very much reduced, have only a slight, hardly perceptible odor. Perret (1913) has commented upon the mild nature of the volcanic eruptive column. He characterizes it as "sweet and clean as the air itself." Comparative analyses of these two types of emission would give some measure of the degree of dilution but, unfortunately, it has not yet been feasible to collect a satisfactory sample of the crater vent vapors. The qualitative difference in the composition of the two emanations, however, is readily apparent to anyone visiting both vents.

A further indication of the essential independence of the quantity of crater vapors and the lava emission is the total lack of any apparent correlation of lava and crater activity. Day by day, during the period of the Taqui flow, from January 1944 to July 1945, the flow of lava appeared relatively constant, the lava issued constantly and at an apparently uniform rate, with quiet exhalation of gases and vapors. The flows, at their point of extravasation, moved without pause or changes, sometimes for weeks at a time. Yet the crater emission was extremely variable; very heavy eruptive columns alternated with reduced activity or even calm, from hour to hour, in an entirely unpredictable manner.

The close relationship of groundwater to the formation of an eruptive column is suggested by the behavior of Kilauea during the eruption of 1924. At that time the lava column of Halemaumau pit sank to a point below the groundwater level, thereby allowing groundwaters access to the lava conduit, producing violent emissions of gases as an imposing eruptive column, in contrast to the normally quiet Kilauean activity (Jaggar 1947).

An ascending lava column, rising from considerable depths to the surface, must inevitably encounter the water table. If the penetrated rocks are, in part, fractured and vesiculated basalt flows, as is the case of Parícutin volcano, a considerable reservoir of water supply is available. The large gushing springs that issue from the basaltic lavas at Uruapan 20 kilometers to the east, and which initiate the Cupatitzio River, indicate some measure of the amount of water which may be available. Adequate groundwater supplies are, indeed, certainly present.

Studies on slag quenching suggest the possibility that a magma can augment its water content by solution of groundwater, which can later be released at the second boiling point (Sosman 1947). Such a secondary enrichment of water in the rising magma is suggested by the strikingly greater fumarolic activity of the first, or Quitzocho, flow than in the later flows.

It seems reasonably evident, therefore, that a volcanic apparatus can set in motion and activate a large quantity of groundwater, perhaps a much larger quantity than its own magmatic emanation. The activation of groundwaters by a volcanic apparatus or by an intrusive body into the zone of groundwaters would furnish a means of extensive rock alteration. This idea has already been suggested by Lawson (1914*a*, 1914*b*) as a possible explanation for the emplacement of certain types of ore deposits. Lindgren (1935) concedes the probability of commingling of meteoric and magmatic waters in the deposition of ores, but would limit it to the zone of groundwaters, or a maximum depth of 10,000 feet. Activated groundwaters may make some contribution to the metallic content of ore deposits, but this addition is believed to be, in general, small. The mechanism does, however, supply an agency for the hydration of larger bodies of rocks than magmatic emanations seems to offer, and to which much of the altered, but unmineralized wall rock may be ascribed.

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STUDIES OF URANIUM MINERALS (V): PHOSPHURANYLITE*

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Abstract

The ill-defined uranium mineral phosphuranylite, hitherto known only from North Carolina on the basis of the original description by Genth (1879) and a later optical description by Larsen (1921), is redescribed from twelve new localities. Phosphuranylite is isostructural with dewindtite, $Pb_3(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$, and its ideal formula is presumed to be $Ca_3(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$. Analyses show much less calcium than required by this formula. It occurs as dense, golden-yellow films composed of microscopic rectangular plates, and is found principally in the weathered zone of pegmatites containing uraninite. Crystallization tetragonal or orthorhombic. Optically biaxial negative with somewhat variable indices: $n\mathrm{X}=1.660$ to 1.690 (colorless to pale yellow), $n\mathrm{Y}=1.696$ to 1.724 (golden yellow), $n\mathrm{Z}=1.698$ to 1.724 (golden yellow); $2\mathrm{V}=0°$ to 35°; $r>v$ strong. Lead substitutes for calcium in some varieties with accompanying increase in indices of refraction.

Introduction

The name phosphuranylite was given by F. A. Genth in 1879 to an ill-defined uranium mineral found in the Flat Rock pegmatite in Mitchell County, North Carolina. The original description stated only that the mineral occurred as microscopic rectangular scales with a deep lemon-yellow color. The formula $(UO_2)_3(PO_4)_2.6H_2O$ was given to the mineral on the basis of an analysis apparently of impure material. These data alone are inadequate to enable the mineral to be recognized with certainty. Fortunately, E. S. Larsen in the first edition of his optical determinative tables (1921, p. 119) cites diagnostic optical data obtained on an authentic specimen borrowed by him. It may be remarked that mineralogists generally are indebted to Professor Larsen for his care in insuring that the voluminous new optical data cited in his Tables were obtained from authentic material.

During the study of an extensive suite of uranium minerals in the Harvard collection, a distinctive and common mineral was recognized which proved to be a hydrated phosphate of uranium and calcium that differed entirely from autunite and meta-autunite, the only other known minerals qualitatively of this composition. This mineral was tentatively identified as phosphuranylite on the basis of Larsen's optical data. *X*-ray study indicated that it was isostructural with dewindtite. Genth's type specimens of phosphuranylite then were borrowed from the part of his personal collection preserved at Pennsylvania State College, and another authentic specimen, labelled as a gift from Genth, was borrowed from

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* Contribution from the Department of Mineralogy and Petrography, Harvard University, No. 316.

Yale University. The latter specimen was that examined by Larsen. The writer wishes to thank Dr. T. F. Bates of Pennsylvania State College and Professor Horace Winchell of Yale University for their courtesy in lending this material for study. Study of this material verified the identification of the mineral at hand as phosphuranylite. A redescription of this species follows.

PHOSPHURANYLITE

Occurrence. Phosphuranylite is a relatively common although inconspicuous mineral that typically occurs in the weathered zone of granite pegmatites. A brief description of the material from the 13 known localities, 12 here described as new, is given in Table 1. It is often associated with autunite or meta-autunite. Less common associates are uranotile, becquerelite, curite, parsonsite and fluorescent hyalite opal. Very commonly the mineral coats irregular cracks in quartz, mica or feldspar in the near vicinity of altered uraninite. In several instances, phosphuranylite was observed as an alteration product of meta-autunite, but it generally is formed earlier than that species. The phosphate content and probably also the calcium content of these species in their pegmatite occurrences appear to be derived principally by the alteration of apatite by meteoric waters in the zone of weathering. The phosphatic solutions thus formed react with uraninite or with intermediate alteration products thereof, such as hydrated oxides, with the formation of the uranium phosphates. Both phosphuranylite and meta-autunite are particularly abundant at the Ruggles pegmatite, Grafton Center, New Hampshire, where apatite occurs in large amounts as a primary mineral immediately associated with uraninite. The uranium silicates uranotile and beta-uranotile, both frequently observed as alteration products or pseudomorphs of uraninite in pegmatites, apparently form when apatite is lacking. The formation of the uranium silicates from uraninite also is preceded by the formation of oxidized-uraninite and of hydrated uranium oxides.

Physical and Optical Properties; Symmetry. Phosphuranylite occurs as thin films or coatings that appear dense, earthy or minutely scaly to the unaided eye, and that are seen to be composed of thin scales with a rectangular outline under the microscope. Macroscopic crystals have not been observed. The color varies from deep golden yellow to rich yellow and is one of the distinctive characters of the substance. The hardness is about 2. The specific gravity, like the hardness difficult to measure because of the mode of aggregation, is about 3.2.

Under the microscope, the mineral appears as irregular or rectangular scales or plates with a perfect cleavage parallel to the flattening. Some

TABLE 1. LOCALITIES FOR PHOSPHURANYLITE

Flat Rock mine, Mitchell County, North Carolina. The Penn State and Yale specimens, and additional specimens from other collections later identified by comparison, consist of smoky or milky quartz, feldspar, muscovite and garnet coated by films of phosphuranylite and meta-autunite. One specimen showed cubical molds of uraninite crystals partly filled with phosphuranylite.

Newry, Oxford County, Maine. A large book of muscovite containing quartz and uraninite along the margins and coated with phosphuranylite along fracture surfaces and in opened cleavages of the mica.

Ruggles mine, Grafton Center, Grafton County, New Hampshire. Films and coatings in fractures and cavities in quartz and feldspar usually near the gummite pseudomorphs after uraninite for which the locality is noted. Found with meta-autunite, parsonsite and a number of other secondary uranium minerals. The indices of refraction vary somewhat in different specimens, with $nZ=1.707$ to 1.716.

Palermo mine, North Groton, Grafton County, New Hampshire. Coatings on aggregates of quartz, muscovite, and kaolinized feldspar containing gummite pseudomorphs. With meta-autunite.

Bedford Hills, Westchester County, New York. Minutely crystalline films on feldspar and quartz, with fluorescent hyalite.

Branchville, Fairfield County, Connecticut. Films and veinlets in altered uraninite. The indices of refraction are relatively high (Table 2) and this material, like that from Urgeirica, may represent a plumbian variety.

Pamplonita, Santander do Norte, Colombia. Earthy films on massive quartz, with meta-autunite. Other specimens from this locality show uraninite altering to curite, schoepite, becquerelite and unidentified minerals.

Carrasca mine, Sabugal, Portugal. Veinlets of flaky phosphuranylite cutting a slab of altered granite whose outer surfaces are thickly incrusted with meta-autunite. The phosphuranylite is admixed with altered (?) meta-autunite of exactly the same specific gravity.

Rosmaneira, Sabugal, Portugal. A slab of deeply altered granite (?) thickly coated with flaky phosphuranylite. The specimen, from the Bello collection, originally was labelled carnotite.

Urgeirica, Cannas de Senhorim, Portugal. A mass of limonitic quartz with drusy cavities containing citron-yellow to brownish yellow microscopic laths and needles mostly clustered in fishbone fashion on crystals of a related mineral. The rectangular laths are beveled by modifying faces. Optically, a zonal growth or mixture of one or several phases with varying indices of refraction. nZ, in part, 1.71–1.72, but mostly 1.72–1.73 and, in part, up to 1.75. All parts of the mixture gave the same *x*-ray powder pattern. The mixture gave a good test for lead, and probably represents a zoning or intergrowth of plumbian phosphuranylite with dewindtite.

Memoes, near Equador, Rio Grande do Norte, Brazil. Crystalline to earthy films on quartz with meta-autunite and unidentified uranium minerals.

Wölsendorf, Bavaria, Germany. A mass of purple-black fluorite, hematite and quartz with fine-grained crusts of a golden-yellow mineral labelled dewindtite. This represents the dewindtite mentioned by Schoep and Scholz in 1931 as occurring at Wölsendorf. The optical properties, somewhat variable, indicate a mixture or zoning of dewindtite and plumbian phosphuranylite.

TABLE 2. OPTICAL PROPERTIES OF PHOSPHURANYLITE

Pleochroic with X (or E) colorless to pale yellow, Y and Z (or O) golden yellow. Values for $nX \pm .005$, for nY and $nZ \pm .003$.

Locality	Optical Character	2V	nX or nE	nY	nZ or nO
Flat Rock mine, North Carolina (Larsen, 1921)	Biaxial—	very small	1.691	1.720	1.720
Flat Rock mine (new data)	Biaxial—	5° to 20°	1.690	1.718	1.718
Mitchell County, North Carolina	Uniaxial—	0°	1.68		1.710
Newry, Maine	Biaxial—, some uniax.—	0° to 5°	1.658	1.699	1.699
Ruggles mine, New Hampshire	Biaxial—	10° to 25°	1.668	1.710	1.710
Palermo mine, New Hampshire			~1.66		1.708
Bedford Hills, New York					1.723
Branchville, Connecticut			1.695		~1.77
Pamplonita, Colombia	Biaxial—	5° to 10°	1.674	~1.724	1.724
Carrasca, Portugal	Biaxial—	5° to 20°	1.669	1.710	1.710
Rosmaneira, Portugal	Biaxial—	15° to 30°	1.660	1.700	1.701
Urgeirica, Portugal	Biaxial—	12° to 22°			1.71 to 1.75
Memoes, Brazil	Biaxial—	10° to 35°	1.660	1.696 to 1.698	1.698
Wölsendorf, Bavaria	Biaxial—	5° to 20°			1.73 to 1.768

material is too fine-grained for satisfactory optical study. The optical character is variable. The mineral rarely is uniaxial negative and ordinarily is biaxial negative with Z always perpendicular to the flattening. 2V is variable, even in material from the same locality, and ranges up to about 35° but usually is small. The biaxial material always shows parallel extinction when crystal outlines are present, and exhibits strong dispersion of the optic axes with $r>v$. The pleochroism is strong, with X (or E) colorless to pale yellow and Y and Z (or O) golden yellow. The indices of refraction are given in Table 2.

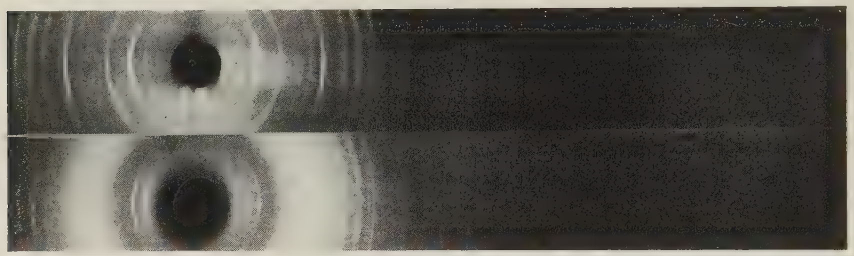

FIG. 1. *X*-ray powder photographs taken in filtered copper radiation of (above) Phosphuranylite from Flat Rock mine, North Carolina, and (below) Dewindtite from Katanga, Belgian Congo.

The occasional uniaxial character and rectangular habit suggests that phosphuranylite may be tetragonal in crystallization. The isostructural mineral dewindtite occurs as microscopic rectangular scales and has been classed as orthorhombic on the basis of the observed biaxial character. The biaxial character of these minerals, however, may be anomalous and is not necessarily indicative of orthorhombic crystallization. This feature may be due to variation in the water content of the crystals. Both autunite and meta-autunite and their analogues are sometimes biaxial with variable 2V due to variation in content of zeolitic water within the limits of stability of the phase, although these species have a tetragonal structure as shown by *x*-ray study (Beintema, 1937, p. 155). The symmetry of phosphuranylite and dewindtite may be tentatively described as tetragonal or orthorhombic, pending *x*-ray study.

The material from Urgeirica, Branchville and Wölsendorf (see Table 2) has relatively high indices of refraction and apparently represents a lead-containing variety in a series toward dewindtite. The latter mineral was described (Schoep, 1925), as biaxial positive with 2E large, $r<v$, and $nX=1.762$, $nY=1.763$. A re-examination here of material from Katanga gave the following results: biaxial negative with moderate 2V and $r>v$; $nX=1.760$ (colorless), $nY=1.768$ (golden yellow), $nZ=1.770$ (golden yellow); Z perpendicular to the plane of flattening.

X-ray Powder Pattern, Systematic Relations. The *x*-ray powder pattern of phosphuranylite is practically identical with that of dewindtite (Fig. 1) The pattern differs entirely from those of the autunite and meta-autunite groups and from all other uranium phosphates and arsenates with the possible exception of renardite and troegerite. Authentic specimens of the two latter minerals were not available for comparison; most specimens labelled troegerite so far examined in detail have proved to be saléeite, uranospinite or hydrogen-uranospinite. The powder spacing data for phosphuranylite and dewindtite are given in Table 3.

Table 3. *X*-Ray Powder Spacing Data for Phosphuranylite and Dewindtite. Copper Radiation, Nickel Filter (in Å)

The measured values of *d* for the two minerals can be taken as identical for all practical purposes. The data here given were measured on phosphuranylite. Different specimens of this mineral show slight differences in detail, notably in a blurring together of lines 11 and 12, 13 and 14, 15 and 16, and in the appearance of very faint, variable extra lines in the region from lines 5 to 18. Line 2 may appear as a doublet due to absorption. Lines beyond no. 49 are omitted.

No.	*d*	*I*	No.	*d*	*I*	No.	*d*	*I*	No.	*d*	*I*
1	10.163	1	13	3.440	1	25	2.156	1	37	1.585	1
2	7.830	10	14	3.363	6	26	2.099	½	38	1.537	2
3	7.196	½	15	3.121	6	27	2.080	2	39	1.506	½
4	6.326	1	16	3.069	4	28	2.040	1	40	1.433	1
5	5.829	8	17	2.940	2	29	2.002	1	41	1.375	1
6	5.372	½	18	2.858	6	30	1.895	5	42	1.358	½
7	4.955	1	19	2.706	1	31	1.852	1	43	1.304	1
8	4.720	2	20	2.585	1	32	1.828	1	44	1.282	1
9	4.332	3	21	2.455	1	33	1.771	1	45	1.252	1
10	4.250	½	22	2.430	2	34	1.719	3	46	1.236	½
11	3.969	9	23	2.243	½	35	1.672	½	47	1.209	½
12	3.834	1	24	2.212	1	36	1.656	1	48	1.197	½
									49	1.179	1

Phosphuranylite contains calcium and ordinarily no lead. Its formula should be analogous to that of dewindtite:

Phosphuranylite $Ca_3(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$
Dewindtite $Pb_3(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$

Partial analyses of phosphuranylite cited in the following section, however, do not agree with this interpretation.

Chemical Composition. Genth's original analysis of the North Carolina mineral reported 4.40 per cent PbO and CaO apparently was not sought. The lead was deducted by Genth as due to admixture of cerussite, although the actual presence of cerussite was not shown, and the formula of the mineral was derived as $(UO_2)_3(PO_4)_2.6H_2O$. A qualitative chemical re-examination of Genth's material shows that lead is absent and that the mineral is in fact a hydrated phosphate of uranium and calcium. Sufficient material was not available for a quantitative analysis of the type material but partial analyses were obtained of specimens from other localities. The results of these analyses, cited in Table 4, are not very consistent and, further, depart somewhat from the ratios expected on the basis of the isostructural relation to dewindtite. Analysis 3, probably the best of those available, approximates to the formula:

Table 4. Chemical Analyses of Phosphuranylite

	1	2	3	4	5	6	7
CaO		3.40	2.3	2.6	2.19	2.74	8.02
PbO	4.40	tr.	tr.	tr.			
UO_3	71.73	72.25	76.4		78.26	70.72	68.15
P_2O_5	11.30	10.55	11.1	10.0	11.10	14.05	13.53
H_2O	10.48	[12.85]	[8.4]		8.45	12.49	10.30
Insol.		0.95	1.8				
Total	97.91	[100.00]	[100.0]		100.00	100.00	100.00

1. Flat Rock mine, North Carolina. Genth, 1879.
2. Carrasca, Portugal. Hallowell, 1949. Contains about 10 per cent of admixed meta-autunite.
3. Rosmaneira, Portugal. Hallowell, 1949. Analysis on 200 mg.
4. Ruggles mine, New Hampshire. Hallowell, 1949. Analyses on 35 mg.
5. $Ca(UO_2)_7(PO_4)_4(OH)_4 \cdot 10H_2O$.
6. $Ca(UO_2)_5(PO_4)_4(H_2O)_4 \cdot 10H_2O$ or $CaH_4(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$.
7. $Ca_3(UO_2)_5(PO_4)_4(OH)_4 \cdot 10H_2O$.

$$Ca(UO_2)_7(PO_4)_4(OH)_4.10H_2O.$$

The most interesting feature of the analyses is the large deficiency of Ca from that expected on the basis of the relation to dewindtite (column 7, Table 4). Possibly the structure of phosphuranylite is like that of dewindtite but with some of the Ca positions vacant, such as occurs in autunite. The latter substance forms an apparently complete series from the ideal composition $Ca(UO_2)_2(PO_4)_2.10H_2O$ to hydrogen-autunite, $H_2(UO_2)_2(PO_4)_2.10H_2O$, either by base-exchange of crystals immersed in acid solutions or by direct crystallization from Ca-deficient acid solutions. The analyses of phosphuranylite, however, do not support this interpretation (column 6, Table 4) and the exact composition of the mineral remains problematic.

Synthesis. Efforts to synthesize phosphuranylite by reaction of water solutions containing uranyl sulfate or acetate, phosphoric acid or trisodium phosphate, and calcium chloride or acetate have proven unsuccessful. The reactions were carried out both at room temperature and at 90°. The precipitates obtained proved in all instances to be phases that resembled meta-autunite.

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PARAGENESIS OF THE RHODOLITE DEPOSIT, MASONS MOUNTAIN, NORTH CAROLINA

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ABSTRACT

A coarse gneissic rock, which occurs on Masons Mountain, Macon County, North Carolina, and consisted originally of rhodolite and hypersthene, was partly transformed, by metasomatism, into anthophyllite-biotite gneiss. Later solutions, which developed quartz and sillimanite with minor rutile and graphite, accompanied pegmatites that contain quartz, feldspar, biotite, kyanite, staurolite, muscovite, and graphite. Both kyanite and staurolite are apparently pyrogenic in origin. The mineralization appears to have been localized along a fault.

INTRODUCTION

The rhodolite garnet localities in Macon County, North Carolina, are about six airline miles north of Franklin, along the flanks of Masons Mountain, a nearly east-west ridge on the east side of the Little Tennessee River. The ridge, which is heavily timbered and covered by brush, has few outcrops, and a deeply weathered mantle obscures the country rock. Most of the rocks underlying the surrounding area are rather uniform in petrology, consisting chiefly of two types, hornblende gneiss (Roan formation) and biotite gneiss (Carolina formation). Kyanitic gneisses are present locally, and corundum gneiss is known from the headwaters of nearby Cowee Creek. Numerous muscovite pegmatites that transect these rocks are related to large intrusives of Whiteside granodiorite-quartz diorite that occur to the southeast in the Highlands-Cashiers area.

The rhodolite variety of garnet was first discovered about 1883 in the gravels of Masons Branch and Cowee Creek together with a rich and unusual suite of detrital minerals: quartz, pyrope, corundum, pleonaste, gahnite, hypersthene, cordierite, kyanite, sillimanite, biotite, hornblende, staurolite, rutile, ilmenite, chromite, zircon, gold, pyrite, chalcopyrite, nickeliferous pyrrhotite, sphalerite, sperrylite, monazite, cyrtolite, and anthophyllite (Hidden and Pratt 1898*A* and 1898*B*, Judd and Hidden 1899, and Henderson, 1931). In the Cowee Creek gravels rhodolite has been found perched on corundum crystals (Judd and Hidden, 1899, 326).

At least three occurrences of the rhodolite in place are known. The first, described by Hidden and Pratt (1898*B*, 468) is one-half mile north of the placer workings near the summit of Masons Mountain. A ledge-like outcrop of rhodolite-biotite rock contains abundant sulfides and traces of sperrylite. A second nearby occurrence was explored by Burnham S. Colburn of Biltmore, North Carolina, in the early 1930s and has been described by Henderson (1931). Here a dike-like mass of rhodolite-anthophyllite (called gedrite) -hypersthene-biotite rock was exposed over a width of 15–20 feet. The anthophyllite from this locality has been restudied by Rabbitt (1948).

Neither of these localities appears to be the one examined by the writer in the spring of 1948, for the rocks found contain abundant sillimanite in addition to the many of the above constituents. This locality is near the west end of the ridge, just east of a newly opened pegmatite quarry (Shepherd vermiculite mine). Pratt (1933) also states that rhodolite was found in gravels "and in the gneissic rocks in several places on this mountain." The writer is indebted to M. V. Denny of the Department of Mineralogy, University of Michigan, for several photographs and to the Department itself for defraying the cost of the thin sections required in this study. E. P. Henderson, of the U. S. National Museum, kindly reviewed the manuscript.

The workings consist of two cuts, of which only the eastern exposes hard rock. This opening, which is ellipsoidal in plan and has a narrow entryway at the southwest end, is 75 feet long in a N. 45°E. direction, 40 feet across, and 60 feet deep at the northeast face. Much of the upper part is in decomposed rock, but fresh gneiss is exposed near the bottom. The gneissic foliation as exposed on the center of the southeast wall strikes N. 65° E. and dips steeply southeast, but in the entryway it strikes N. 70° W. and dips 80° NE. Thus it appears that the mineralized zone has been developed along either a fault or along the fractured axis of a sharply folded syncline that plunges steeply eastward. In view of the continuance of the zone southwestward, it seems more likely that a fault has been the localizing structure. Henderson (1931, 564) states for the Colburn locality " . . . the rhodolite-gedrite rock has been intruded along a fault where there has probably been some displacement." The second cut 30×20×15 feet in size, adjoins the larger opening at the southwest end; however, no fresh rock is exposed in it.

RHODOLITE ROCKS AND ASSOCIATED ROCKS

The rock types exposed in the cut may be grouped together as follows:

1. Pegmatite and quartz vein material.
2. Rhodolite rocks.
    - (*a*) Rhodolite-quartz rock with accessory graphite.
    - (*b*) Biotite-rhodolite-quartz gneiss with variable amounts of hypersthene and anthophyllite.
    - (*c*) Biotite-rhodolite-sillimanite gneiss with variable amounts of quartz and anthophyllite.
3. Anthophyllite-rhodolite-quartz gneiss with variable amounts of sillimanite and biotite.

4. Anthophyllite-quartz gneiss.
5. Biotite-anthophyllite gneiss with variable amounts of sillimanite.

These rocks are only the main types present and represent a somewhat arbitrary grouping, for the varieties intergrade over short distances. The texture is commonly coarsely gneissic and the foliation is due largely to mineralogical banding, although rocks rich in anthophyllite (type 4) are somewhat schistose in structure, with well oriented amphibole blades. The outstanding features of the exposures are the mineralogical variability and the heterogeneous distribution of the rock varieties. Pods, lenses, and irregular segregations of the various rock types are common, and contacts between units are gradational.

The most conspicuous rocks are those containing abundant rhodolite. Coarse rhodolite-quartz aggregates occur as ragged, pod-like masses, as much as 6–8 inches long, in somewhat finer-grained biotite-rhodolite-quartz gneiss. Some of the pods contain as much as 60–70% garnet, in a subordinate interstitial network of quartz. The rhodolite shows its distinctive rose-lavender color and is clear but fractured. Irregularly rounded grains, $\frac{3}{4}$ inch or less in diameter, were observed. According to Pratt (1933) the largest piece of gem rhodolite ever found weighed $43\frac{1}{2}$ carats, and the largest cut stone weighed $13\frac{3}{8}$ carats. Between the garnet grains and the quartz are brilliant, $\frac{1}{8}$ inch spangles of graphite. Clumps of dark reddish brown biotite flakes and a few ragged crystals of olive green hypersthene, one-half inch long, also occur in the segregations.

The matrix in which the pods are set is somewhat less coarse-grained, contains less garnet, and ranges from strongly gneissic to granoblastic in texture. It varies in composition from biotite-garnet gneiss with hypersthene, quartz, and graphite through biotite-anthophyllite-garnet gneiss, to anthophyllite-biotite gneiss with minor garnet. Some of the anthophyllite occurs as blades more or less normal to the contacts of the garnet-rich pods. Sillimanite becomes abundant locally as tufts and sinuous films of white fibers, $\frac{1}{2}$ inch long.

Thin sections of these rocks yield detailed textural evidence on the mineral paragensis. Garnet, colorless in the thin slices, forms large ragged poikiloblastic grains enclosing quartz, biotite, and anthophyllite. Quartz is the most abundant of the inclusions, which are commonly confined to central parts of the garnet grains. This quartz, in contrast to the more abundant late quartz, shows no strain or flamboyant structure. Minute seams of chlorite replace garnet.

Several types of veins cut broken garnets. Anthophyllite appears in monomineralic veins as the earliest of the transecting minerals. The pale golden brown amphibole, which commonly forms coarse subhedral prismatic blades, is unoriented to rather well-oriented. Rocks in which anthophyllite predominates contain garnet only as very small, strongly corroded specks.

A second generation of veins contains biotite, quartz, and a few ragged anthophyllite relicts. In general biotite is later than anthophyllite, for the mica follows grain boundaries of the amphibole and corrodes adjoining blades. It also replaces anthophyllite along cleavages. Most of the quartz of the veinlets crystallized after biotite. This is shown by the restriction of biotite to the margins of some veins, i.e., against garnet, whereas quartz occupies the central parts. The marginal mica flakes have feathery contacts against the quartz.

Both the anthophyllite and biotite veins in garnets are abruptly transected, in some places at right angles, by veinlets and streaks of sillimanite or of quartz and sillimanite. Disseminated sillimanite prefers to replace biotite. In the banded gneisses sillimanite is confined to the biotite layers and does not occur in anthophyllite-rich parts. It forms fringes of small fibers around the margins of biotite flakes or lies along biotite-anthophyllite grain contacts. Aggregates of sillimanite prisms contain relict shreds of biotite. Sillimanite commonly forms euhedral crystals or crystal groups, coarse-grained and poorly oriented. Most of it is associated with abundant quartz in quartz-sillimanite bands and lenses in which coarse euhedral grains of rutile are locally abundant. Rutile also forms vein-like structures across and within garnet and along anthophyllite cleavages and grain margins.

Other quartz-sillimanite bands contain parallel graphite flakes as well as rutile, and some graphite also has been developed along anthophyllite cleavages. The quartz of these layers usually shows conspicuous flamboyant structure, wavy extinction, a good orientation and minutely sutured contacts. It resembles vein quartz that has been somewhat metamorphosed. At least some of the sillimanite is later than quartz, for "trains" of small sillimanite crystals follow zones of fracturing in granulated quartz. In general, rocks that contain abundant quartz also contain much sillimanite but very little biotite.

Rounded zircons occur in garnet, anthophyllite, biotite, and quartz. Hypersthene is closely associated with coarse garnet and forms large ragged grains replaced by biotite and anthophyllite.

All these highly variable rock types are cut by veinlets of quartz, sillimanite, and quartz-sillimanite. Some of the quartz veins are three inches thick and contain specks of magnetite. Fibers of white sillimanite, $\frac{1}{4}$ inch long, also are plastered against the sides of fractures cutting across the gneissic structure.

Near the entryway, a pegmatite zone, four feet thick, consists of three stringers that coalesce upward into a two-foot sill along the foliation of the gneisses. Mottled blue to white kyanite blades, 0.1×0.3 inch in size, are scattered throughout the central parts of the sill. Quartz, kaolinized feldspar, and biotite flakes partly altered to vermiculite also are present. The kyanite blades are randomly oriented; a few are curved. Locally clusters of tightly interlocking, generally parallel blades form fist-sized masses.

From the above relationships it is apparent that the rhodolite-bearing rocks and their associates have undergone several transformations. Amongst the earliest minerals that remain are rhodolite and hypersthene. The unstrained quartz and the zircon probably antedate some of the garnet, for they appear as inclusions in the central parts of the rhodolite. Much anthophyllite appears to have formed at the expense of garnet. It veins that mineral, and anthophyllite-rich rocks contain only small rhodolite relicts. The chemical compositions of the rhodolite and anthophyllite are very similar (Henderson, 1931), except that the anthophyllite has a lower FeO content. It seems likely that rhodolite and hypersthene were metasomatically transformed into anthophyllite and biotite. This change requires chiefly the subtraction of iron and silicon and the addition of hydroxyl and aluminum.

The second stage was marked by the introduction of solutions that became increasingly rich in silicon and replaced much of the earlier rock substance with quartz, minor rutile, and a trace of graphite. During the final phases aluminum was introduced and sillimanite was formed, mainly at the expense of biotite. These later stages coincided with the intrusion of the kyanite pegmatites and the formation of the quartz veins.

Kyanite-Staurolite Pegmatite

About 200 feet southwest of the rhodolite occurrence a new bulldozer cut (Shepherd mine) has been dug to mine vermiculite. The cut, which is 200×300 feet in plan and about 40 feet deep at the face, is largely in decomposed biotite gneiss. Near the bottom of the face a zone of pegmatite dikes trends N. 30° W. and dips 45° NE. Eight dikes, ranging in thickness from one to four feet, are exposed over a 20-foot width. The country rock, a dark fine-grained biotite gneiss containing a few grains of pink garnet, has been converted to a granoblastic aggregate of fine-grained vermiculite, around and between the pegmatites.

Exposures of some of the pegmatites show no kyanite, but other dikes are locally rich in the mineral. The chief constituents are clear to milky quartz and kaolinized white feldspar. Pegmatites with abundant kyanite contain more quartz than feldspar. Some dikes have small core pods of massive white quartz or of "burr rock"—massive quartz studded with

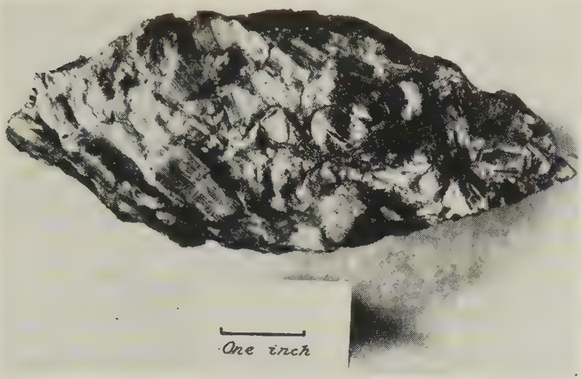

FIG. 1. Kyanite pegmatite from Shepherd mine, Masons Mountain, North Carolina. The warped strip in the right central part of the specimen is biotite. Other blades are kyanite in quartz-kaolinized feldspar matrix.

small, subparallel muscovite books. Biotite plates several inches across are not uncommon. Locally a dark reddish brown mineral occurs rather abundantly in pods as long as two inches. In the field it was thought to be garnet, but microscopic examination shows it is staurolite. Further search of the specimens revealed a perfect staurolite crystal, one-half inch across with the typical flattened, eight-sided cross section and brilliant prism faces. Reexamination of kyanite pegmatite from the rhodolite cut (described above) also disclosed several minute staurolite grains. Indices and pleochroism of the staurolite are:

$\alpha=1.739$, pale yellow
$\beta=1.744$, pale orange
$\gamma=1.749$, golden orange

Most of the staurolite is unusually free of inclusions or alteration, but a few grains show typical quartz-inclusion structure.

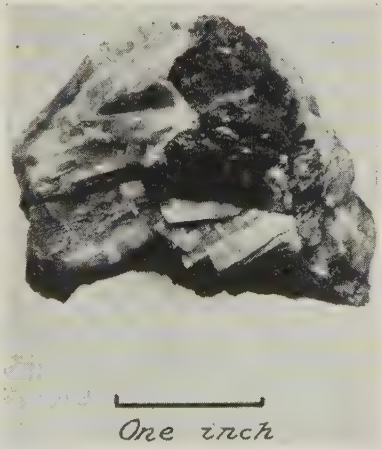

FIG. 2. Anhedral staurolite (dark) molded between kyanite blades, Shepherd pegamite, Masons Mountain, North Carolina.

Kyanite, in blades as much as three inches long and an inch wide, varies from pale blue to deep blue. Some of the smaller blades show a tendency toward subparallel orientation, but the larger are randomly arranged (Fig. 1). Others are curved and broken. Kyanite developed after biotite, for some blades bisect the large mica plates. Staurolite, which crystallized after kyanite, is molded against the latter (Fig. 2) and sends off thin stringers into kyanite blades. Some staurolite has been granulated.

Small spangles of graphite are disseminated through parts of the pegmatites. Scattered vugs, $\frac{1}{4}$ inch across, are stained by manganese and iron oxides and may have formed through destruction of some of the kaolinized feldspar.

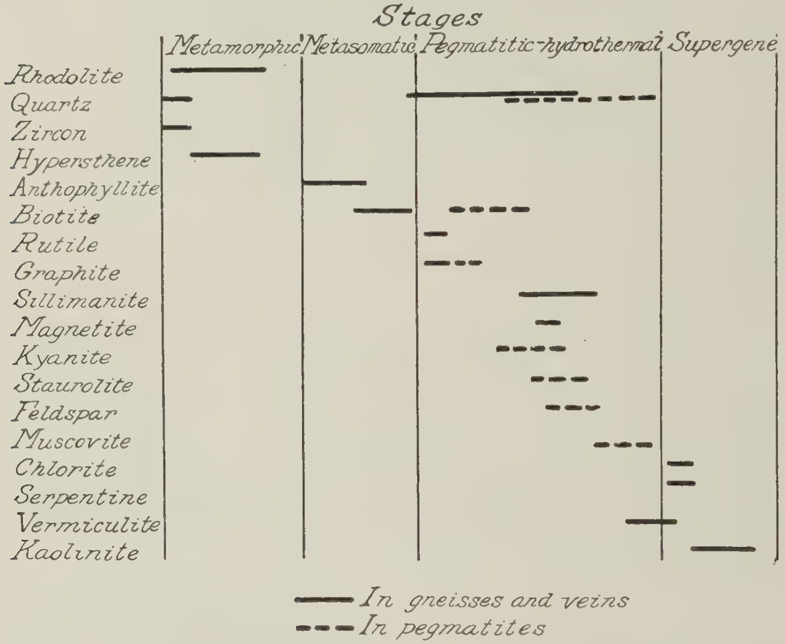

FIG. 3. Schematic diagram showing paragenetic sequence of the minerals of the rhodolite deposits and associated pegmatites, Masons Mountain, North Carolina.

CONCLUSION

This deposit represents one of the few recorded occurrence of coarse staurolite as an important pegmatitic constituent. A few other examples of staurolite in pegmatites are known (for example, Laubman and Steinmetz, 1920, Scholz, 1925 and Frondel, 1940), but the mineral appears to be either xenocrystic or metamorphic in origin. The writer has also recently found staurolite in kyanite-quartz veins in a kyanite-staurolite schist in the pre-Beltian Cherry Creek Series along Cherry Creek, Madison County, Montana. This occurrence may be analogous to that cited by Chapman (1946), who describes quartz veins, some with staurolite and others with staurolite concentrations in the immediate wall rock, in staurolite schist of the Littleton formation in west-central New Hampshire.

If any country rock material was assimilated by the Shepherd pegmatites, the process took place at depth before the pegmatite magma was intruded up to its position of crystallization, for the wall rocks are staurolite-free. There is no doubt that these dikes represent the southwestern continuation of the pegmatite zone exposed in the entryway to the rhodolite cut. Probably the aluminous pegmatites and the solutions that produced the quartz-sillimanite mineralization had their origins in a common source. It is interesting to note that kyanite crystallized in the pegmatites and sillimanite formed in the wall rocks, yet conditions of pressure and temperature could not have been radically different. The complete paragenetic sequence of the minerals in the gneisses as well as those in the pegmatites is shown in figure 3.

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PETROGENESIS OF KATUNGITE AND ITS ASSOCIATES

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Abstract

The young volcanic fields near Ruwenzori are characterised by ultrabasic potassic volcanic rocks made up of various combinations of melilite and/or augite; kalsilite and/or leucite and/or potash-rich glass; and, with few exceptions, olivine; invariable accessories are perovskite and black ores. The types include *katungite* (olivine+melilite+potash-rich glass), *mafurite* (olivine+augite+kalsilite), and *ugandite* (olivine+augite+leucite) and all possible transitions are represented. The activity was intensely explosive. The pyroclasts invariably contain fragments both of basement rocks and of a cognate sub-volcanic suite composed of pyroxene and/or biotite and/or olivine (O.B.P. series). The rocks and minerals of the latter appear simultaneously with the volcanic types as fragments in the tuffs and inclusions in the lapilli, etc. It is shown that the series proto-katungite (olivine-free)—katungite—alnöite—olivine-ouachitite is a genetic one that could have been formed from O.B.P. by magmatic material even more ultrabasic than proto-katungite. In an earlier hypothesis this magmatic material was regarded as "emanations" (E) of alk-aluminous and cafemic composition, but the sources of the "emanations" and of the O.B.P. rocks were then left unexplained. It is now suggested that the alk-aluminous material (and all the SiO_2 of the resulting rocks) came from granite (G) and its derivatives, and that the cafemic material came from "magmatic carbonatite" (C). For katungite (K) the hypothesis can be expressed in the form $C+G=O.B.P.+K$. Additional potash, (e.g. in mafurite) is accounted for by the demonstrated fact that biotite formed at greater depths became altered and gave up part of its potash to the liquid phase. Brief reference is made to the possible application of the "magmatic carbonatite" hypothesis (*a*) to the petrogenesis of soda-rich alkali rocks; and (*b*) to the long-continued uplift of the rift valley regions of Africa and the origin of the rifts themselves.

Dedication

Nearly twenty years ago, during a memorable visit to Harvard, I first had the privilege of discussing with Professor Larsen some of the interests we share in common, including the petrogenesis of biotite-pyroxenites and melilite-rocks. Since then I have turned again and again to his stimulating publications on the Iron Hill (Uncompaghre) complex for guidance in my efforts to solve the related problems raised by the pyroclasts and lavas of the volcanic fields near Ruwenzori. I have long suspected that the most significant clue to their origin lay in the carbonatite core of the Uncompaghre complex—and, of course, in the analogous occurrences of carbonatites in so many other complexes of alkali rocks. But only recently have I found a way to follow up the clue, as I hope, successfully. It therefore seems appropriate to select this new development as the subject of my contribution to the *Festschrift* organised in honour of Professor Larsen. If, in advocating the active role played by carbonatite regarded as a magma, instead of considering carbonatite as the material assimilated by some other magma, I have, so to speak, turned inside out

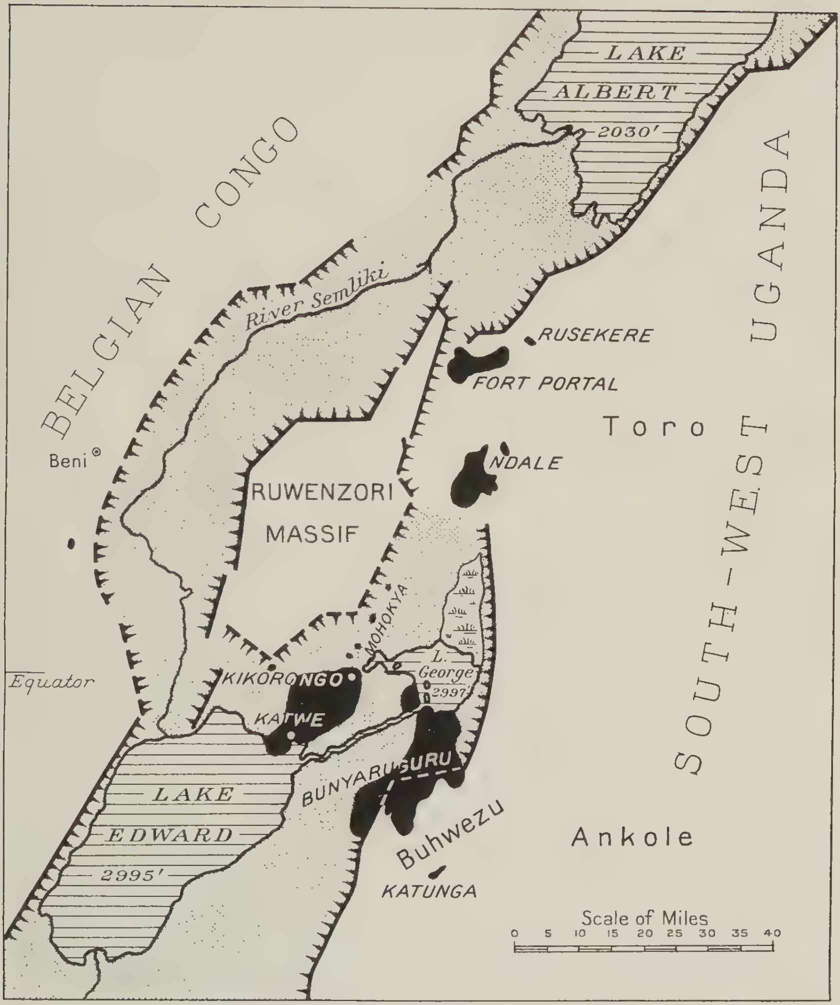

FIG. 1. Sketch map of part of the Western Rift Valley of Africa showing the horst of Ruwenzori and (in black) the neighbouring volcanic fields. The Bukangara field is indicated by the black spot above the 'OR' of KIKORONGO.

the hypothesis favoured by Professor Larsen, he himself will be the last to object on any grounds other than those based on factual evidence. It is a genuine pleasure to dedicate to him this token of friendship and admiration.

THE VOLCANIC FIELDS NEAR RUWENZORI AND THEIR MATERIALS

Half encircling the imposing horst of Ruwenzori, in and adjoining the

Western Rift of Central Africa, there are several areas of tuffs and explosion craters accompanied, in some of the areas, by lava flows or near-surface plugs and sheets. The various fields (Fig. 1) are distinguished under the following names: (*a*) *Katunga*, in the plateau country in the south of the province; (*b*) *Bunyaruguru*, lying athwart the eastern wall of the rift depression; (*c*) *Katwe-Kikorongo*, in the rift, between L. Edward and L. George; (*d*) *Bukangara*, on the slopes leading up to the southern foot of Ruwenzori (possibly an extension of (*c*) to the north-west); (*e*) an isolated *unnamed* explosion crater lying well to the west of Ruwenzori on the Butembo Road, 14 miles south of Beni in the Belgian Congo; (*f*) a line of craters on the *Mohokya* bench extending N.N.E. from Kikorongo towards Ndale; (*g*) *Ndale*, where the low wall of the L. George depression dies out to the north; (*h*) *Fort Portal*, where the eastern wall of the L. Albert rift dies out to the south; and (*j*) *Rusekere*, on the plateau to the north-east.

Although each of these areas has its own local characteristics, the group as a whole constitutes a highly distinctive petrographic province, the leading features of which were first recorded in a preliminary contribution (Holmes and Harwood, 1932) based mainly on the reconnaissance collections made in the early days of the Uganda Survey. Subsequently my old friend and collaborator, the late Mr. A. D. Combe, began a systematic survey of the province which he completed only a few months before his lamented death in 1949 (see Holmes, 1949). At intervals the specimens collected by him, amounting in all to well over a thousand, have been sent to me for petrological study. It is expected that a joint memoir—which will be a lasting monument to Combe's insight as a vulcanologist—will eventually be published (see Combe, 1930–1944).

Apart from the Pleistocene lacustrine and fluviatile deposits of the Kaiso Series, which occupied much of the floor of the rift depression, the underlying rocks are of Pre-Cambrian age, the sequence being:

Post-K.A. Granites
Karagwe-Ankolean (K.A.) System: mainly argillites, phyllites and quartzites

Toro System: mainly schists, quartzites, amphibolites, gneisses and granites.

The upper part of the Kaiso Series is of Middle Pleistocene age. In the Bunyaruguru and Katwe-Kikorongo fields it grades upwards into subaqueous volcanic tuffs which are themselves perforated by a succession of later vents. Some of the younger crater-rims are so little affected by erosion that they cannot be very old. Moreover, there are native traditions of volcanic activity (Wayland, 1934), some of which are preserved

TABLE 1. CHEMICAL ANALYSIS OF MEMBERS OF THE O.B.P. SERIES

	1	2	3	4	5	6	7
SiO_2	50.39	43.35	44.39	33.99	38.18	42.59	47.33
Al_2O_3	3.04	9.67	6.37	5.86	16.52	3.82	6.84
Fe_2O_3	2.61	5.26	5.11	9.35	2.90	2.55	.02
FeO	5.42	2.65	4.86	8.54	5.24	4.96	5.62
MgO	13.14	15.74	14.20	7.53	19.32	33.84	19.31
CaO	22.57	12.20	17.02	18.23	1.16	2.84	16.99
Na_2O	.56	.44	.50	.69	.56	.26	.41
K_2O	.08	4.74	2.35	2.38	9.01	2.58	.88
H_2O+	.14	.59	.35	.96	1.22	2.96	.55
H_2O-	none	.07	.05	.18	.58	.63	.21
CO_2	none	.04	.04	.30	none	.25	.02
TiO_2	1.96	4.38	3.91	8.65	5.18	1.60	1.26
ZrO_2	—	none	—	none	—	—	none
P_2O_5	tr.	tr.	.26	2.80	—	.22	.21
Cl	—	tr.	.04	.02	—	.05	tr.
F	—	.11	.04	.09	—	.12	.01
S	—	tr.	.02	.03	—	.18	.01
Cr_2O_3	—	.05	.03	none	.14	.23	.32
V_2O_3	—	.04	.05	.07	—	.01	.04
NiO	—	.04	.01	none	—	.21	.04
MnO	.05	.09	.10	.15	tr.	.10	.12
BaO	tr.	.30	.16	.13	.16	.13	.03
SrO	.12	none	.03	.05	.22	.02	none
Li_2O	—	tr.	tr.	none	—	tr.	tr.
	100.08	99.79	99.89	100.00	100.17	100.15	100.22
Less O	—	.04	.03	.05	—	.11	.01
	100.08	99.75	99.86	99.95	100.17	100.04	100.21

1. Pyroxenite. Xenolith from ejected block of olivine-rich ugandite, C.4035, Kakunyu crater, Bunyaruguru (Combe and Holmes, 1945, p. 377). *Analyst:* W. H. Herdsman.
2. Biotite-pyroxenite. Ejected block, C.2786, vent on S.S.W. end of Lutale Ridge, Bufumbira (Holmes and Harwood, 1937, p. 29). *Analyst:* H. F. Harwood. Total includes CuO = .03.
3. Biotite-pyroxenite. Ejected block, G.20, Katwe crater (Holmes and Harwood, 1937, p. 30). *Analyst:* H. F. Harwood.
4. Sphene-rich biotite-pyroxenite. Ejected block, K.4, Katwe crater (Holmes and Harwood, 1937, p. 30). *Analyst:* H. F. Harwood.
5. Glimmerite. Xenolith from ejected block of olivine-rich ugandite, C.4034, Kakunyu crater, Bunyaruguru (Combe and Holmes, 1945, p. 377). *Analyst:* W. H. Herdsman.
6. Biotite-peridotite. Xenolith from ejected block of olivine-rich ugandite, C.4034, Kakunyu crater, Bunyaruguru (Holmes, 1945, p. 377). *Analyst:* H. F. Harwood.
7. Biotite-bearing augite-peridotite. Ejected block, C.1963, Mabungo crater, Bufumbira (Holmes and Harwood, 1937, p. 23). *Analyst:* H. F. Harwood.

in the names of the craters: e.g. Kasesankaranga (in the Fort Portal area) means "spewer of roasted material." The volcanic activity thus seems to have continued intermittently from late Middle Pleistocene to comparatively recent times.

The tuffs and other pyroclasts contain fragmental materials consisting of:

(1) The underlying rocks and their minerals.
(2) The rocks and minerals of a sub-volcanic suite of biotite-pyroxenite and related types, including pyroxenite, glimmerite and peridotites. The suite will be referred to collectively as the "O.B.P." series, after the dominant minerals olivine, biotite and pyroxene. Common but variably developed minerals in the biotite-pyroxenites are melanite, ilmenite, sphene, apatite and calcite, usually replacing the earlier minerals in the order stated. Perovskite and potash nepheline are also occasionally present. The rocks range in grain size from extremely coarse to very fine and are often patchy, exhibiting sudden changes of texture and composition. The types and their associations can be closely matched by the corresponding rocks described from the Libby Stock (Larsen and Pardee, 1929) and the Iron Hill complex (Larsen, 1942). Representative samples of the O.B.P. series have been analysed and the results are listed in Table 1, together with those for two examples from the Bufumbira field (North Kivu) to the south.
(3) Lapilli, bombs and ejected blocks of *katungite* and other ultrabasic potash-rich volcanic types, the assemblage of which for the whole province is schematically summarised in Fig. 2. Apart from xenocrysts, not a trace of feldspar or hornblende

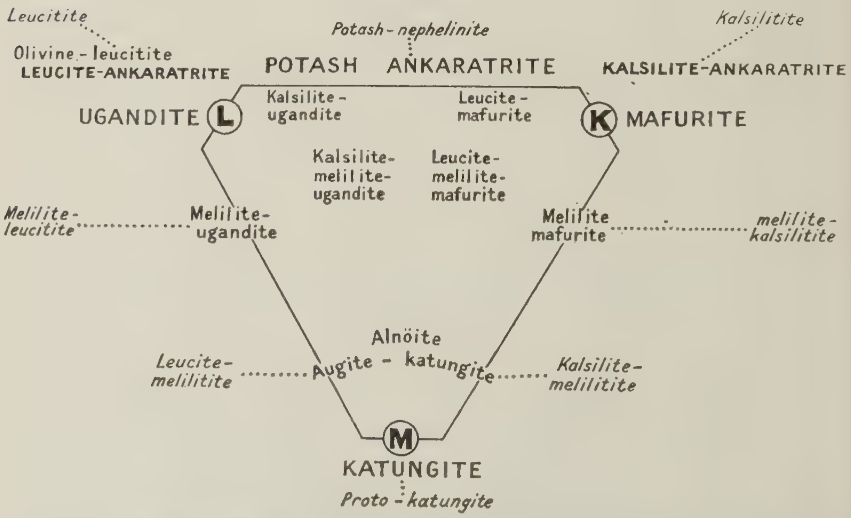

FIG. 2. Triangular classification of the rock-types occurring in the volcanic fields near Ruwenzori. L = Leucite and augite; K = Kalsilite and augite; M = Melilite and potash-rich glass, occasionally with kalsilite or leucite. All the types contain olivine, except those named in italics. Biotite varieties (including heteromorphs such as olivine-ouachitite) of some of the types are known, but are not named except in the case of alnöite. The ankaratrite series differs from the mafurite-ugandite series in being conspicuously rich in augite and relatively poor in olivine.

has been detected in any of the volcanic rocks examined. The critical minerals of the suite are melilite and/or augite, and kalsilite and/or leucite, together with olivine in all the types except those—mostly very rare—named in italics in the outer part of Fig. 2; biotite occurs in some varieties and accessories common to all are perovskite and opaque ores, often accompanied by apatite. In katungite, however, which is ubiquitous and generally by far the most abundant type, the potash is mainly present in a glassy base from which kalsilite, leucite or biotite have crystallised in only a few of the many occurrences studied; phillipsite, however, is commonly present in veinlets and as a lining to vesicles. Because of the paucity of kalsilite and leucite, katungite is placed at one of the corners, M, of Fig. 2, the others being L, occupied by *ugandite* or *leucite-ankaratrite* (melanocratic types of olivine-leucitite which are respectively rich in olivine and augite); and K, occupied by *mafurite* or *kalsilite-ankaratrite*, the corresponding types with kalsilite in place of leucite. The diagram itself makes clear the composition of the many transitional varieties.

Lava flows are rare, except in the Fort Portal area, where highly carbonated melilite-nepheline-leucitite occurs as numerous small outpourings (Holmes and Harwood, 1932, p. 379). Flows of katungite are known only from Katunga itself (Combe, 1937). In Bunyaruguru, mafurite, the rock in which kalsilite was first discovered (Bannister and Hey, 1942; Holmes, 1942), occurs in a lava sheet truncated by the wall of a crater adjoining Lake Mafuru. In the same field there are other flows and a lava-dome or plug consisting of mafurite and related types (Combe and Holmes, 1945), and also flows of ugandite and melilite-ugandite—from Kazimiro—not yet described. The Katwe-Kikorongo field is noteworthy for the abundance of various members of the potash-ankaratrite series, mostly occurring as ejected blocks, but also as lava-sheets and plugs in the craters of Mbuga and Nyabugando.

One of the most significant features of the province is the fact that in the earliest tuffs minerals of the O.B.P. series and lapilli of katungite (see Table 2, 8) invariably appear together. Since an overwhelming proportion of the lapilli contain xenocrysts of augite and biotite, and sometimes of olivine, all obviously derived from the O.B.P. series, it is clear that the latter crystallised before the consolidation of the associated katungite. In this sense the oldest O.B.P. rocks are older than the oldest katungite, but it is also clear that the O.B.P. series and the associated katungite magma originated contemporaneously. The same relationships occur throughout the province. The lapilli, bombs and blocks of katungite—and also those of all the other volcanic types, including the flows—contain xenocrysts, and often aggregates, of the O.B.P. minerals. While katungite contains mainly biotite and augite, ugandite carries a higher proportion of olivine and potash-ankaratrite of augite; mafurite contains variable amounts of all three minerals, often with a conspicuous abundance of biotite, much of which has been severely altered, as described on p. 789. There is, of course, no reason to suppose that all the O.B.P.

rocks and minerals are older than all the volcanic types. The two sets of materials may have continued to form contemporaneously throughout the period of activity. No standard sequence of volcanic types can be established, but in many of the crater walls a record of several eruptions of katungite can be recognised, with intervening appearances of members of the ugandite, potash-ankaratrite or mafurite series.

THE KATUNGITE SERIES: PETROGENETIC CONSIDERATIONS

The rock now known as katungite was originally described as melilite-basalt or olivine-melilitite. Detailed investigation of the specimens from the two Katunga flows showed, however, that they differ from olivine-melilite in being both potash-rich and free from augite; hence a new name was felt to be justified. As usual, leucite can only rarely be detected, but kalsilite has now been found in a number of the specimens. Both minerals are restricted to the lower parts of the flows, where the lavas are richer in potash than nearer their sources. Natrolite is a common zeolite near the source but lower down phillipsite increases in amount and natrolite dies out. The later parts of the lava to be extruded were evidently progressively poorer in potash and relatively richer in soda. Confirmation is provided by the following analyses (W. H. Herdsman), listed in order of time of emergence of the analysed materials:

		K_2O	Na_2O	K_2O/Na_2O
C.4418	Bomb, outer face of crater	4.21	1.34	3.2
C.4407	Lava, near end of western flow*	4.09	1.32	3.1
C.4411	Lava, 50 feet below source of flow	3.84	2.91	1.3
C.4412	Lava, near source of flow	3.55	2.96	1.2

* For complete analysis see Table 2, 9.

This evidence illustrates a characteristic feature of the rocks of the province—the compositional variability of the volcanic rocks—observed not only in the lava flows, but even in single ejected blocks. It also exemplifies a general tendency, shared with the Birunga or North Kivu field, for the later products of an eruption to be relatively less potassic than the earlier. It seems likely that soda was partly removed from the upper part of the magma column, possibly as carbonate or some other volatile compound. Oldonyo l'Engai (south of Lake Natron in the Eastern Rift) erupted soda-rich vapours, mainly carbonate, in 1917, as recorded by Hobley (1918). That an immense concentration of gas had risen to the head of the magma column at Katunga before the outbreak occurred may be inferred from the visible indications of a powerful explosive phase. The vent, 800 feet in diameter, was drilled through Toro schists and a vast accumulation of blocks and slabs of these rocks form the cone and the crater rim. Tuffs with katungite lapilli came next, followed in turn by bombs and finally, extruded from the slopes of the cone, by the eastern and western flows.

Kalsilite-katungite has now been recognised in the tuffs from a dozen different localities in Bunyaruguru. Only one band has been found, however, in which all the lapilli consist of this type and so could be separated for analysis (Table 2, 10). The analysed type-material has been already figured and described (Holmes, 1942; Combe and Holmes, 1945). In other occurrences it has been noticed that wherever xenocrysts derived from quartz-bearing basement rocks are present—as they often are—kalsilite has failed to develop in their vicinity. Similarly in the mafurite-ugandite and potash-ankaratrite series it is found that development of leucite instead of kalsilite (and also of augite instead of melilite) is often demonstrably due to the incorporation of sialic material. It therefore seems probable that kalsilite-katungite would be much more plentiful, were it not for siliceous contamination.

Confirming this inference is the fact that xenocrysts of quartz in the katungite bomb C.4418, referred to above, have been transfused into glass having the composition of a potassic obsidian (for analysis see Holmes, 1936, p. 415). Before parting with Al, K and H in exchange for Si, the molten lava must evidently have been even more ultrabasic and potassic than the resulting solid katungite. Innumerable other examples of the same phenomenon are widespread, but much more spectacular is the discovery that granitic xenoliths and xenocrysts derived from them or from other sialic rocks have been transformed into leucite or leucite glass and eventually, in some cases, into leucitite and olivine-leucitite (Holmes, 1945). So far examples of leucitization have not been found in katungite, but they are by no means rare in members of the mafurite-ugandite series.

The katungite from Chamakumba crater (Table 2, 11) illustrates another common type of transformation which is of great significance. The rock has an extremely dark groundmass and contains two varieties of melilite: one, clear and beautifully fresh, giving inky blue interference colours; the other showing low normal colours and sometimes having a patchy isotropic core. The change from one variety to the other occurs suddenly and crystals can be found which are 'normal' at one end and 'inky blue' at the other. The 'normal' variety has clearly been derived from xenocrysts of augite, for most of these have corroded and embayed margins which have been replaced by this variety of melilite. The rims consist of a coarse felt of melilite laths which are separated from the unaltered augite by a continuous peripheral band of melilite with the same optical properties as the laths. Xenocrysts of biotite have also contributed to the resulting katungite, for they are marginally trans-

TABLE 2. ANALYSES OF KATUNGITE, ALNÖITE AND OLIVINE-OUACHITITE

	8	9	10	11	12	13	14
SiO_2	34.23	35.37	33.52	33.22	35.51	37.93	38.94
Al_2O_3	8.02	6.50	8.04	9.71	6.83	6.59	6.92
Fe_2O_3	6.62	7.23	5.88	6.68	9.68	6.81	5.27
FeO	5.34	5.00	5.50	5.30	2.70	4.37	5.09
MgO	9.92	14.08	13.54	12.12	11.67	14.54	11.58
CaO	16.54	16.79	15.22	15.64	16.00	15.23	15.95
Na_2O	1.20	1.32	1.42	1.51	1.56	.88	1.01
K_2O	3.39	4.09	4.26	3.54	3.30	2.65	3.96
H_2O+	2.80	2.78	2.34	3.28	3.11	3.38	2.26
H_2O-	1.72	1.15	1.68	.80	1.31	1.42	1.20
CO_2	4.02	.09	.96	.42	1.47	.50	2.12
TiO_2	4.56	3.87	6.04	6.08	4.88	4.12	3.88
ZrO_2	—	none	—	—	none	—	—
P_2O_5	.96	.74	.82	1.12	1.18	1.03	.91
Cl	—	.02	—	—	tr.	—	—
F	.14	.16	—	.08	.27	.16	.13
S	.12	.35	—	—	—	—	.14
SO_3	—	—	—	—	.13	.06	—
Cr_2O_3	—	.01	—	—	.02	—	—
V_2O_3	—	.03	—	—	.04	—	—
NiO	—	.19	—	—	.02	.03	.09
MnO	.22	.24	.15	.52	.22	.18	.23
BaO	.20	.25	.15	.17	.27	.30	.18
SrO	.23	.04	.44	.05	.24	.25	.19
Li_2O	—	none	—	—	tr.	—	—
	100.23	100.36	99.96	100.24	100.41	100.51	100.05
Less O	.09	.15	—	.02	.11	.07	.09
	100.14	100.21	99.96	100.22	100.30	100.44	99.96

8. Katungite. Lapilli separated from the first volcanic horizon of the Upper Kaiso Series, C.5945, near M.P.75 on the Fort Portal Road, N.W. edge of the Bunyaruguru field. *Analyst:* W. H. Herdsman.
9. Katungite. Western flow of Katunga volcano, C.4407 (Holmes, 1937, p. 205). *Analyst:* A. W. Groves. Total includes CuO 0.06.
10. Kalsilite-katungite. Lapilli separated from tuff, C.4012, N.W. rim of Changabe crater, Bunyaruguru (Combe and Holmes, 1945, p. 367). *Analyst:* W. H. Herdsman.
11. Katungite. Ejected block, C.3509, Chamakumba crater, Bunyaruguru. *Analyst:* W. H. Herdsman.
12. Katungite. Lapilli from tuff, G.21, Katwe crater. *Analyst:* H. F. Harwood.
13. Alnöite. Ejected block, G.56, Katwe crater. Another sample gave Na_2O 1.07; K_2O 3.63 (Holmes and Harwood, 1932, p. 405). *Analyst:* H. F. Harwood.
14. Olivine-ouachitite. Ejected block, C.5844, from 20 feet below the rim of Katwe crater. *Analyst:* W. H. Herdsman.

formed to a cloud of black-ore specks in an obscure residuum. Alteration of biotite, of which there is ample evidence throughout the area, would obviously provide some of the potential leucite or kalsilite required for katungite, besides some of its olivine and black ores.

These observations indicate that katungite has originated, at least in part, from biotite-pyroxenite, or, more generally (remembering that corroded xenocrysts of olivine are also occasionally present), from the materials of the O.B.P. series. Theoretically, at any rate, katungite could be formed from the O.B.P. series by the desilicating action either of a magma much more ultrabasic than itself or, at the limiting extreme of ultrabasicity, of "emanations" of cafemic (including TiO_2, P_2O_5, etc.) and alk-aluminous composition. If this concept in either form has any basis in fact it should be possible to recognise (*a*) a series of types bridging the gap between katungite and the O.B.P. series; and, in the other direction, (*b*) types having a composition more ultrabasic than normal katungite. Both expectations have been realised.

The series (*a*) is illustrated by katungite, alnöite and olivine-ouachitite from the tuffs and agglomerates of the walls of Katwe crater. The micrometric measurements given below (including two other measured katungites) indicate the range of the melilite proportions:

Rock and Locality			Melilite (Vol. per cent)
Katungite, Katunga,	C.4407	(*An.* 9)	40
Chamakumba,	C.3509	(*An.* 11)	38
Katwe,	G.21	(*An.* 12)	37
Alnöite, Katwe,	C.5775		26
	C.5778		21
	G.56	(*An.* 13)	18
	C.5779		5
Olivine-ouachitite, Katwe,	C.5884	(*An.* 14)	0

Katungite contains abundant melilite and no augite other than sparse xenocrysts and minute amounts formed by reaction against transfused quartz. Olivine-ouachitite has abundant augite and no melilite. The various specimens of alnöite fall between. In these the augite is mostly too fine grained for accurate measurement, but it is clear from inspection that there is a rough complementary relationship between melilite and augite.

The more ultrabasic type (*b*) referred to above is the rock for which the name *proto-katungite* has already been suggested (Holmes, 1942, p. 199; 1945, p. 317). The name implies the supposition that it represents the earliest known stage in the magmatic ancestry of katungite. Somewhat paradoxially—though emphasizing the extreme basicity—proto-katungite differs mineralogically from katungite in being free from olivine. Occurrences of lapilli and ejected blocks of this type turn up sporadically in the tuffs and agglomerates of Bunyaruguru and Katwe-Kikorongo, their most abundant development being in some of the oldest of the pyroclasts. Without exception the rocks are crowded with xenolithic biotite and augite (rarely olivine) and most of them are highly charged with carbonates.

The analysed example (Table 3, 15 *a*) is the freshest part of an eight-inch block of fine-grained gray lava, mottled with minute calcite-lined vesicles and conspicuously sprinkled with xenolithic material. Apart from very rare relics of phyllite and quartzite, the xenolithic material consists of fragments of biotite-pyroxenite and its minerals, the biotite being margined and veined by clouds of black ores and perovskite in an obscure background, while the augite is margined with reaction rims of melilite. Only a trace of xenocrystic olivine is present. In the lava itself, small laths of melilite are very abundant, together with grains, octahedra and clusters of magnetite and golden brown or clove-coloured perovskite, all embedded in a greenish grey to brownish base which is isotropic or feebly cryptocrystalline and copiously dusted with specks of black ore and perovskite.

The bulk composition of the material is that of a typical katungite, from which it follows that in this particular case, which is typical, katungite = proto-katungite + biotite-pyroxenite (+ a very little olivine). A closer approach to the composition of proto-katungite is given in Table 3, 15*b*.

Between proto-katungite and katungite with abundant olivine there is every gradation, olivine-poor varieties being very common. Taking the series as a whole, a rough complementary relationship can often be noticed between the respective abundances of phenocrysts of olivine and xenocrysts of biotite and augite. At one extreme are the holo-magmatic lavas of Katunga with well developed olivine and no xenolithic relics. At the other are the occurrences of proto-katungite, invariably rich in xenolithic relics and free from authigenic olivine. Plotting the analyses of Table 2, and either average biotite-pyroxenite or average O.B.P. (Table 3, 17), on a silica basis, results in an approximately straight-line variation diagram. The proto-katungite analysis, 15*a*, fits well into the series and so, obviously, must 15*b*. The series proto-katungite—katungite—alnöite—olivine-ouachitite—biotite-pyroxenite (or O.B.P. for the more olivine-rich varieties) is therefore a genetic one, in the sense that it can be completely accounted for in terms of its end-members.

In looking through a comprehensive suite of specimens of katungite-bearing tuff, one cannot avoid noticing that sialic basement rocks, often including granite, contribute a very high proportion of the accidental fragments, whereas such debris is rare within the katungite lapilli, the dominant xenolithic material being derived from biotite-pyroxenite. This generalisation is also true of the occurrences of proto-katungite. Moreover, olivine and olivine aggregates from biotite-peridotite are much more abundant in the non-volcanic materials of the tuffs than in the volcanic lapilli. It may therefore be inferred that at the time of explosive eruptivity the katungite and proto-katungite magmas existed in an environment of biotite-pyroxenite, which was probably ensheathed by peridotite or olivine-rich members of the O.B.P. series, the latter being covered, or perhaps in part intercalated with, the sialic rocks above the explosion level. This inference is consistent not only with the conclusions already reached, but also with the possibility later to be discussed that both magma and O.B.P. may in part have been derived from granitic and other sialic rocks. Some support for this, until recently, unfamiliar idea is provided (*a*) by the observation that biotite and augite of O.B.P. types have been found as replacement minerals in a few of the ejected blocks of basement rocks; and (*b*) by the evidence that the biotite-pyroxenite and -peridotite of the Newry Complex (Northern Ireland) represent part of a basic front which developed by replacement of country rocks as a complementary product of underlying granitization (Reynolds, 1944, pp. 234–236).

PETROGENESIS: A SUGGESTED HYPOTHESIS

In the first paper on katungite, being already impressed with the evidence of transfusion and metasomatism brought about by migrating "emanations" (gaseous transfer, or diffusion), I proposed the hypothesis that katungite, K, might be a rheomorphic product of the action of essentially silica-free emanations, E, on biotite-pyroxenite (Holmes, 1937, p. 216). Briefly, and more generally, this idea can be qualitatively expressed in the form

$$\text{K} = \text{O.B.P.} + \text{E.}$$

Graphically, the same idea is represented in Fig. 3, where the point E on the base of the triangle indicates the composition of the emanations required to transform average O.B.P. (the point m) to katungite, the field of which is shown in black. The line Em, prolonged upwards, passes through biotite-pyroxenite, which for all the olivine-poor varieties of

TABLE 3

	Proto-katungite		K. Av. Katungite	Average O.B.P.	G. Av. Granite	C	Alteration of Biotite	
	15*a*	15*b*	16	17	18	19	20	21
SiO_2	33.89	31.91	34.54	42.89	72.51	—	37.05	42.77
Al_2O_3	8.27	8.45	7.77	7.45	14.73	—	15.04	9.02
Fe_2O_3	7.03	7.71	7.17	3.97	.69	11.56	4.12	5.09
FeO	5.21	5.18	4.74	5.33	.79	9.48	5.10	4.18
MgO	10.93	9.48	12.18	17.58	.49	29.18	19.74	13.48
CaO	16.98	17.85	15.94	13.00	1.06	29.61	.33	12.21
Na_2O	1.42	1.62	1.39	.49	2.79	−.85	.23	1.42
K_2O	3.65	3.76	3.70	3.15	5.28	1.55	9.55	5.05
H_2O+	2.08	2.33	2.83	.97	.72	3.59	4.15	1.39
H_2O-	1.19	1.39	1.32	.25	.18	1.68	.30	.82
CO_2	3.27	3.96	1.38	.09	.23	1.57	—	.40
TiO_2	4.43	4.56	5.05	3.85	.20	9.31	2.81	3.61
ZrO_2	—	—	—	none	.02	—	—	—
P_2O_5	.97	1.06	.95	.58	.13	1.54	—	—
Cl	—	—	.01	.02	.03	−tr.	—	—
F	.18	.21	.16	.07	.07	.18	.40	—
S	.14	.16	.19	.05	.03	.25	—	—
Cr_2O_3	—	—	.02	.13	none	.13	—	—
V_2O_3	—	—	.04	.04	tr.	.08	—	—
NiO	—	—	.10	.05	none	.17	—	—
MnO	.26	.29	.27	.09	.03	.38	.28	.15
BaO	.21	.23	.21	.13	.09	.27	1.26	.15
SrO	.29	.34	.20	.06	none	.30	none	.11
Li_2O	—	—	tr.	tr.	tr.	—	—	—
	100.40	100.49	100.16	100.24	100.07	99.98	100.36	99.85
Less O	.11	.13	.16	.05	.04	.15	.17	—
	100.29	100.36	100.00	100.19	100.03	99.83	100.19	99.85

15*a*. Proto-katungite (including xenocrysts: approximately, augite, 12%; biotite, 5%; olivine and black ore, 1%). Ejected block, C.6065, from the rim of one of the oldest craters, Lugazi, Bunyaruguru. *Analyst:* W. H. Herdsman.

15*b*. Proto-katungite. Estimated composition after subtracting 18% average O.B.P. column 17) from the all-over analysis, 15*a*. Subtracting biotite-pyroxenite corresponding to the xenocrysts actually present yields essentially the same result.

16. Average of five analyses (Table 2, 8–12) of katungite.

17. Average of seven analyses (Table 1, 1–7) of the O.B.P. series. Plotted as "*m*" on Fig. 3.

18. Average of six analyses of Post-K.A. granites (King, 1939, p. 138). Plotted as 'Granite' on Fig. 3.

19. Calculated composition of "C," the carbonatite material (less liberated CO_2) re-

katungite would serve as well as average O.B.P. Proto-katungite is represented by the white dot in the katungite field nearest to E and is probably a close approach to the most ultrabasic silicate rock that could exist. The hypothesis would now take the form

$$PK = O.B.P. + E.$$

This hypothesis, however, is no more than a half-way stage, since it leaves both O.B.P. and E unaccounted for. And even if O.B.P. be regarded, as it was in 1937, as primary peridotite metasomatically modified by emanations rich in Ca, Fe, Ti, K, Al, etc., the source of the various emanations still remains unknown. Moreover, the hypothesis fails to account for the mafurite series, except by involving the action of emanations even richer in potash. Despite the obviously close relationship between katungite and mafurite, neither can be derived from the other by addition or subtraction of crystalline phases, whether magmatic or xenolithic.

The combination of high potash with cafemic constituents is the geochemical peculiarity that especially challenges explanation. It would be a purely *ad hoc* assumption to regard primary peridotite or the material of the substratum as a potential source for the potash of either potash-rich rocks or the alk-aluminous emanations necessary to generate such rocks. Supporting this rejection of primary peridotite is the very low radium content of rocks of the latter type (Davis, 1947). Members of the O.B.P. series and the comparable xenoliths of the kimberlite pipes contain 50–280 times as much radium as primary peridotite (Holmes and Paneth, 1936). Basaltic materials cannot be regarded as a promising source for potash and in any case they are out of the question, since they play no part in the province under discussion. The only obvious source of potash would appear to be granitic rocks and their pelitic derivatives: let us say granite, for short. This possibility, however, implies that granite supplied not only potash, but also alumina and silica, and that the cafemic materials must have come from an essentially silica-free source, such as carbonates.

Theoretically, reaction of high-temperature granite magma with carbonate rocks of appropriate composition would solve the problem. But there are fatal objections to this variant of the limestone-assimilation hypothesis. There was no granite magma to incorporate carbonate rocks.

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quired to generate katungite (16) and O.B.P. (17) from granite (18). Iron oxides expressed as total FeO = 19.88.

20. Biotite. Core of xenocryst in kalsilite-ugandite, C.4788, Upper Kabirenge lava, Bunyaruguru (Combe and Holmes, 1945, p. 377). *Analyst:* H. F. Harwood.

21. Alteration product of biotite. Rim of same xenocryst as No. 20 (*Ibid.*). *Analyst:* H. F. Harwood.

Apart from the carbonate deposits representing, e.g. the hot-spring phase of the volcanic activity itself, the only known carbonate rocks are thin intercalations of limestone locally present in the Karagwe-Ankolean System. Blocks of these have been ejected from some of the vents of Bunyaruguru but, unlike the overwhelmingly more abundant sialic rocks, they show no signs of alteration or incorporation. Moreover, they are geochemically of inadequate composition. What is required would have to be a mixture of dolomite, siderite and magnesite (in proportions such as 65:25:10 respectively) with a good deal of TiO_2, H_2O and P_2O_5 and notable amounts of other minor ingredients: a most improbable sediment.

The only remaining alternative which would not be altogether fantastic is to postulate reaction of "magmatic carbonatite" with granitic rocks. This idea has several advantages. The temperature would be adequate for the work to be done. There would be ample supplies of high-pressure gas available for the drilling of vents, for the production of widespread tuffs and the ubiquitous development of drop-like lapilli, and for the transport of "emanations." Known carbonatites in Africa (Dixey, Campbell Smith and Bisset, 1937; Pulfrey, 1944; Davies, 1947; King, 1949), Iron Hill (Larsen, 1942), Fen (Brögger, 1921), Alnö (von Eckermann, 1948) and many other localities are often dolomitic and ferruginous and are geochemically characterised by the appropriate assemblage of other elements necessary to the hypothesis. In this respect the only unusual feature about the Uganda rocks is their very high content of TiO_2, but even this can be matched elsewhere. Since $MgCO_3$* would be the first to give up CO_2 and react with granite, it is only to be expected that the carbonatites actually exposed (whether magmatic or hydrothermal) should be richer in $CaCO_3$ than the original materials. Carbonatites are naturally not to be found in comparatively recent vents, but there is ample evidence in Uganda of a source of carbonate in depth. Many of the tuffs, and especially the older ones, are saturated with $CaCO_3$, which occurs both as cement and as a replacement mineral. The lavas of the Fort Portal and neighbouring areas are so highly carbonated that in many cases they have almost lost their identity. Travertine deposits of hot-spring origin are developed on a considerable scale along the outer slopes and foot of the Mohokyo bench, and pinnacles of travertine have been built up in Katwe and other craters (Combe, 1944, p. 14).

The hypothesis is now proposed that magmatic carbonatite, presumably derived from the substratum, ascended through fractures into the sialic layer where it reacted with, say, granite, G, to form crystalline O.B.P. and molten proto-katungite or katungite, K, with eventual loss of most of the liberated CO_2. Qualitatively we can write

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* Probably $FeCO_3$ would give up CO_2 even earlier than $MgCO_3$.

$$\mathrm{C + G = O.P.B. + K}$$

where C represents the composition of the carbonatite, less all the liberated CO_2 except that fixed in katungite and O.B.P. It is easy to show that this hypothesis is only a genetic development of the old one, qualitatively expressed as

$$\mathrm{K = O.B.P. + E.}$$

Let us now make the symbols quantitative by using italics. Suppose that an amount $(C+G)$ of the carbonatite and granite yields an amount $O.B.P.$ of crystalline phases and an amount E of residual material, i.e.

$$(C + G) = O.B.P. + E. \tag{1}$$

In the limiting case, where all the SiO_2 of G has been used up, E becomes the "emanations" of the old hypothesis (cf. Fig. 3). Let E react with an amount $a\ O.B.P.$ to produce an amount K of katungite. Then

$$K = a\ O.P.B. + E \tag{2}$$

which expresses the old hypothesis. Eliminating E from (1) and (2), we have

$$(C + G) = O.B.P. + K - a\ O.B.P. = (1 - a)\ O.B.P. + K \tag{3}$$

which expressed the new hypothesis.

It is, of course, impossible to say to what extent the two stages (1) and (2) represent the real facts of the very complex processes that must actually have been involved. On Fig. 3 the line CG represents all possible mixtures of C and G, and m represents average O.B.P. The residual material could theoretically lie anywhere on the part of the line mE below its intersection with the line CG. Katungite lies on this part of the line, proto-katungite being nearer E and alnöite being nearer m. Olivine-ouachitite falls on the O.B.P. side of the intersection, this representing the rare case where the proportion of O.B.P. to E has been higher than in equation (2)

It is interesting to notice that E (on Fig. 3) also represents the composition of the "emanations" required to transform granite into leucitite. The base-line of the triangle (Fig. 3) represents all the possible compositions of "emanations" from the alk-aluminous material of the corner A to the essentially cafemic material of the corner C. The material required to transform granite into leucite is represented by a point that closely approaches A.

It is now possible to estimate, at least roughly, the composition of C in equation (3). Adopting the figures in analyses 16, 17 and 18, we can

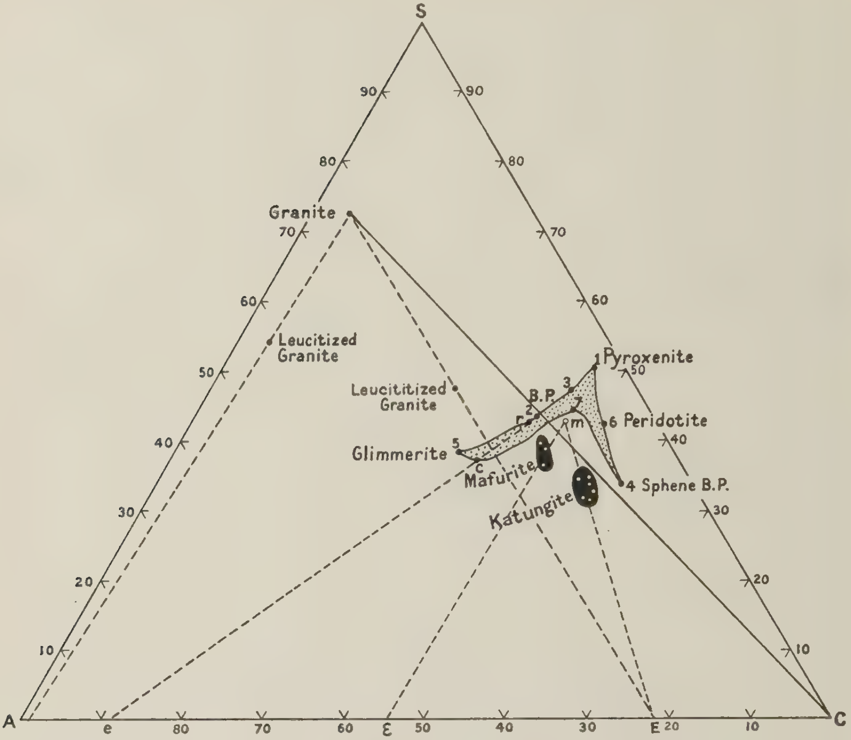

FIG. 3. Triangular diagram to illustrate the chemical relationships of some of the chief rocks under discussion. S = SiO_2; A = $Al_2O_3 + K_2O$ and Na_2O; C = the sum of all the remaining constituents (mainly cafemic oxides). m = average composition of the O.B.P. group (dotted), the numbers referring to the analyses of Table 1. c and r = the core and rim of an altered biotite xenocryst (Table 2, 20 and 21). e represents the composition of the material liberated by the alteration of biotite from c to r; E that of the material required to transform m into katungite; and ϵ that of the material required to transform m into mafurite.

write for any given constituent

$$x\,\mathrm{K} + (100 - x)\ \mathrm{O.B.P.} = y\,\mathrm{G} + (100 - y)\ \mathrm{C}. \tag{4}$$

SiO_2 and Al_2O_3 give two equations which can be solved for x and y, assuming that these two constituents are ideally absent from C. Inserting the values so found, equation (4) can now be written.

$$60.9\ \mathrm{K} + 39.1\ \mathrm{O.B.P.} = 51.9\ \mathrm{G} + 48.1\ \mathrm{C}. \tag{5}$$

Solving for the remaining constituents, the composition of C is found to be as listed in Table 3, 19. If the same procedure is carried out for proto-katungite, 15*b*, equation (5) becomes

$$39.5 \text{ PK} + 60.5 \text{ O.B.P.} = 53.2 \text{ G} + 46.8 \text{ C}$$

indicating that much less PK has been generated, as would be expected. The composition of C in this case is nearly the same as before.

Inspection of column 19 in Table 3 shows that C is essentially cafemic oxides, with high TiO_2 and considerable H_2O and P_2O_5, together with notable amounts of all the characteristic minor constituents of the province. Na_2O is negative, implying a loss which is probably to be referred to gaseous transport, as already suggested. The presence of K_2O is less easily disposed of, since it suggests that granite may not be capable of producing all the K_2O required. This would not disturb von Eckermann, however, who writes, in postulating a carbonate magma to account for the rocks of the Alnö complex (1948, p. 161): "The magmatic melt, or rather liquid, is suggested to have been a high tensioned one . . . , consisting of mainly carbonates, predominatingly potassic. . . ." It is not surprising, therefore, that he also writes (p. 157): "No explanation, however, has been offered as to the origin of this liquid and none will be given in this memoir." Since my own main purpose in this paper is particularly to try to account for the potash of the rocks under discussion, the appearance of even a little K_2O in the composition of C calls for further discussion. This is the more necessary because mafurite contains up to 7% K_2O(Holmes, 1942) and the composition of C required to generate this type of lava would have to be correspondingly richer in K_2O.

Fortunately this excess potash can be readily accounted for without assuming abnormally potash-rich granites and without arbitrarily taking it for granted in C. Reference has already been made to the marginal alteration of the biotite relics in katungite. More conspicuous relics, with dull stony rims surrounding fresh cores, are especially common in the members of the mafurite-ugandite series. Analyses of an investigated example from kalsilite-ugandite show that the composition of the rim (analysis 21) approximates to biotite-pyroxenite and that the biotite (analysis 20) lost mainly Al_2O_3, MgO, K_2O and H_2O during its alteration. Obviously, biotite formed at greater depths and higher pressures, gave up part of its potash, etc. to the liquid phase, and so provided the material required for mafurite. In Fig. 3 the composition of the rim is represented by r (near B.P.) and that of the core by c (near Glimmerite). The line through r c meets the base of the triangle at e, which represents the liberated material. A combination of e and C, or of e and E, gives ϵ, representing the "emanations" required to generate mafurite from average O.B.P. The slight excess of K_2O required for katungite is, of course, similarly accounted for.

Lack of space forbids discussion of the many side issues that arise from this preliminary outline of the hypothesis here proposed. It should, however, be kept in mind that for simplicity of exposition only the limiting silica-free compositions of carbonatite and emanations have been considered. The actual compositions need not have been entirely silica-free. The liquid carbonatite may, for example, have been associated with normal periodotite magma, in which case it would only be necessary to add the constituents of peridotite in equal amounts to each side of equation (4). This would, of course, not affect the arithmetical solution of the problem. It should also be remembered that various compositions could be used for O.B.P. in place of the average here adopted. The sequence of crystallisation and metasomatism and the effects of migration of volatiles are other features that call for subsequent discussion.

Before concluding, however, two other points deserve brief mention. The first is that the same kind of hypothesis, suitably modified, should be applicable to the soda-rich alkali rocks of Africa, particularly on the eastern side, many of which are associated with plug-like carbonatites (see King, 1949). Here the difficulty is to get rid of the potash which has been the special problem of the Western Rift. The only conjecture that can be offered is that the reactions may have occurred at a high crustal level in the Western Rift (and other potassic provinces), but at a much lower level in the east (and other sodic provinces). The granitic rocks of the lower levels of the "granite layer" are likely to be richer in soda than potash, not only because granodiorite is more abundant than potassic granite, but also because granitization seems everywhere to have involved a late upward migration of potash. Moreover, reaction with the materials of the "basaltic layer" is likely to be common and there again soda is strongly favoured. Another probability is that biotite formed in depth would remain stable and so would drain off potash from the primary reacting materials. The residual magma and volatiles that ascended to higher levels would thus become increasingly enriched in soda. It is hoped that a satisfactory extension of the hypothesis to embrace a wide variety of alkali complexes may eventually be developed along these lines.

The second point refers to the origin of the rift valleys. One of the most striking features of many parts of Africa, including the rift-valley regions, is the cumulative uplift that has taken place since at least the early Cretaceous. How is this to be accounted for? From an isostatic point of view it can only mean that changes have taken place in the substratum whereby the density of the material of the latter, especially near the base of the crust, was reduced. Thermal expansion might do a little in this way, but it would surely be far from sufficient to meet the requirements. The only remaining possibility would seem to be an appropriate change of composition. Ascent of carbonate-carrying material, and its gradual concentration beneath the crust, would not only provide for the eventual development of carbonatites and alkali rocks in the parts of the crust now exposed to view; it would also inevitably lead to the uplift of the overlying country. Rift valleys may well have been formed, in the manner experimentally investigated by Cloos (1939), where the upper parts of the substratum became particularly enriched in carbonate material. If so, the negative anomalies of gravity could be readily accounted for, without recourse to the otherwise unsatisfactory compression hypothesis. The cause of the great uplift of Ruwenzori would also cease to be a tantalising mystery. The "magmatic carbonatite" hypothesis thus promises to co-ordinate a variety of enigmatic phenomena which have hitherto defied explanation.

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CHILDRENITE—EOSPHORITE SERIES*

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Abstract

An optical study of the childrenite-eosphorite series shows twinning in all well-crystallized members indicating monoclinic symmetry. In singly terminated crystals twinning is on {100}; in doubly terminated crystals twinning is on both {100} and {001}. The dimensions of the pseudo-orthorhombic unit cell as determined on crystals from Newry, Maine are: $\alpha_0=10.45$ Å, $b_0=13.49$ Å, $c_0=6.93$ Å, $\beta=90°00'$. A re-examination of material from old localities and study of crystals from three new localities shows a fairly uniform increase in refractive index and specific gravity with increase in FeO. The optical properties of zoned crystals indicate a complete isomorphous series between the iron and manganese end-members. The optical orientation and dispersion for eosphorite is: $Z\wedge c=3°-8°$, $r<v$; for high iron childrenite, $Y\wedge c=6°-8°$, $r>v$. Extrapolating from measured values of specific gravity, G=3.20 for the iron end-member, G=3.05 for the manganese end-member.

Introduction

Eosphorite was first described by Brush and Dana (1878) from Branchville, Connecticut. They pointed out the crystallographic similarity of their new mineral to childrenite and showed that the formulas of the two were identical with the exception of the dominant divalent metal which was manganese in eosphorite and iron in childrenite. Later Drugman (1914) extended the knowledge of this isomorphous series by crystallographic measurements and chemical analyses of childrenite from the Crinnis mine, Cornwall, and of eosphorite from Poland, Maine. Other analyses of eosphorite are given by Landes (1925) and Palache and Shannon (1928), and of childrenite by Otto (1935).

During the optical examination of crystals of eosphorite in a recently acquired specimen from Newry, Maine, the writer noticed the presence of twinning. An examination of specimens from other localities showed this phenomenon to be found in all well-crystallized material. Since twinning in eosophorite has not been previously reported, a reexamination of the whole group was felt to be desirable.

Eosphorite from three new localities, Red Hill and Black Mountain, both in Rumford, Maine, and North Groton, New Hampshire, was available for study.

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* Contribution from the Department of Mineralogy and Petrography, Harvard University, No. 317.

Because of Professor Larsen's outstanding contributions to the knowledge of the optical properties of minerals, it seems fitting to include this optical study in a volume dedicated to him. Moreover, it was he in the various capacities of teacher, employer, colleague and friend who influenced the writer to undertake a career in mineralogy.

EOSPHORITE

The name eosphorite was given by Brush and Dana (1878) to the iron-poor analogue of childrenite. The mineral described by them contained 7.22% FeO (average of three analyses). The formula for the manganese end-member of the isomorphous series can be written $MnAl(PO_4)(OH)_2.H_2O$. Eosphorite was described as orthorhombic with crystallographic measurements and axial ratios closely resembling those of childrenite, $FeAl(PO_4)(OH)_2.H_2O$.

Newry, Maine. Eosphorite from Newry was described by Palache and Shannon (1928). During 1949 several specimens of the mineral were found there which proved to be very similar to the original. The smallest crystals (about 0.1 mm. long) are a light brown but with increasing size they become darker and the largest (4 mm. long) are black. The habit of most of the crystals is the same with a nearly square cross section formed by the pinacoids *a*{100} and *b*{010} as shown in Fig. 1. However, a few of the smaller crystals are flattened on {010} (Fig. 2). All are terminated by *s*{121}.

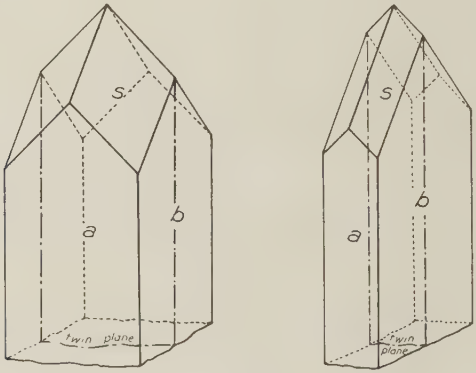

Fig. 1 Fig 2

Eosphorite. Newry, Maine.

Optical examination of these smaller more transparent crystals showed, when lying on {010}, a difference in the extinction positions of the two halves of the crystal. In all cases the line dividing the two parts was sharp and the extinction angle varied from 6° to 8°. There seems to be no other interpretation but that the crystals are twinned by reflection on {100} and that they are monoclinic rather than orthorhombic. Figure 3 is a photomicrograph of such a twin and Fig. 4 shows the optical orientation of the two individuals.

It will be noted in Fig. 3 that in addition to the line marking the twin boundary, there are zone lines parallel to the apex of the crystal. From the indices of refraction it is apparent that these lines represent changes in the composition from the base to the top. The following optical properties were obtained on a crystal one millimeter long:

Optical Properties of Eosphorite, Newry, Maine

Orientation	Base of crystal n(Na)*	Tip of crystal n(Na)	
$X=b$	1.638	1.639	Opt.—
Y	1.660	1.664	2V=50°
$Z\wedge c=6°–8°$	1.667	1.671	$r<v$, strong

From the curves given in Fig. 11, these variations in refractive indices would indicate that the base of the crystal contains approximately 10% FeO and the tip approximately 13% FeO. However, the FeO content in the crystals as a whole is 10.10 (average of three determinations) indicating that the iron-rich portion is small.

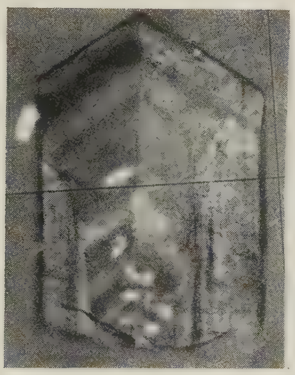

Fig. 3. Photomicrograph of eosphorite lying on {010} showing twinning.

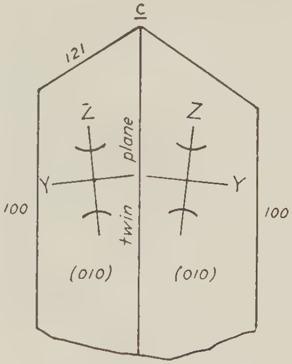

Fig. 4. Optical orientation of twinned individuals shown in Fig. 3.

Shannon, in his analysis of eosphorite from Newry, reported 3.74% FeO. Two new iron determinations** made on the original material of Palache and Shannon (1928) gave 9.56% and 10.28% FeO. The FeO in the new material was found to be 10.45%. Since the optical and other physical properties of the two specimens are identical, it would appear that the original analysis is in error.

Because the Newry eosphorite crystals are rough and striated, goniometric measurements were unsatisfactory and did not lead to the determination of axial ratios. The measurements, however, were sufficiently good to determine that the only forms present are *a*{100}, *b*{010}, and *s*{121}. The pseudoorthorhombic symmetry is produced by the repetition of {121} at the back of the crystal by twinning.

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* The indices of refraction as given throughout were determined by the immersion method using sodium light and are correct to ±0.001.

** All the new chemical analyses reported here were made by F. A. Gonyer, Harvard University.

Rotation photographs and zero, first and second layer line Weissenberg *x*-ray photographs were taken about both [010] and [001], using a single individual. Additional Weissenberg photographs were taken using a twin. The corresponding photographs of the twin and single crystal were identical showing that the crystallographic angle β is 90° and that the twinning brings about no nonparallelism in the lattice. In Table 1 are given the dimensions of the unit cell and the axial ratios for the established pseudo-orthorhombic position, which it seems wise to retain. The corresponding values for childrenite determined by Barnes (1949) are also given.

TABLE 1. DIMENSION OF THE UNIT CELL AND AXIAL RATIOS OF EOSPHORITE AND CHILDRENITE

	Eosphorite (Newry)	Childrenite (Tavistock)
	Unit Cell Dimensions	
	$a_0=10.45$ Å	$a_0=10.38$ Å
	$b_0=13.49$ Å	$b_0=13.36$ Å
	$c_0=6.93$ Å	$c_0=6.911$ Å
	Axial Ratios	
Morphology:	$a:b:c=0.7745:1:0.5139$*	$a:b:c=0.7780:1:0.5258$
Unit Cell:	$a_0:b_0:c_0=0.7745:1:0.5137$	$a_0:b_0:c_0=0.7766:1:0.5173$

* Morphological ratios from the unpublished manuscript of the 7th edition of Dana's *System of Mineralogy.*

The extinctions observed on the Weissenberg photographs of eosphorite are the same as those reported for childrenite by Barnes (1949),who determined the space group as *Bba*2. It seems impossible to reconcile these extinction criteria with the optical properties. However, inasmuch as the optical evidence for monoclinic symmetry is strong, the lower symmetry is here assumed to be correct and the mineral monoclinic.

The specific gravity of eosphorite from Newry was given by Palache and Shannon as 3.069. Several new determinations using the large black crystals gave various values ranging from 3.06 to 3.08. However, the specific gravity of the smaller transparent crystals was determined as 3.10 by suspension in methylene iodide. Since these crystals lack the cracks and slight alteration present in the larger ones, the higher value is assumed to be more nearly correct. Using the cell dimension for eosphorite given above, the calculated specific gravity is 3.112.

Buckfield, Maine. Eosphorite from Buckfield was described by Landes (1925); his original material was available for study. The eosphorite is in long yellow prismatic crystals as shown in Fig. 5. The chemical analysis made by Vassar (Landes, 1925) shows 1.38% FeO, the lowest percentage of iron reported in any eosphorite. The indices of refraction are also the lowest and probably are very near those of the iron free end member. The specific gravity is 3.10. The crystals are all twinned by reflection on {100} and each of the individuals shows an extinction angle against the twin plane of 5°–6°. Most of the crystals are zoned in much the same way as those from Newry and Red Hill with the material with higher refractive index at the tips. The major portion of each crystal has nearly constant refractive indices, those given by Landes.

Optical Properties of Eosphorite, Buckfield, Maine

Orientation	Bulk of crystals n(Na)	Tips of crystals n(Na)	Opt.—
$X=b$	1.629	1.637	$2V=45°–50°$
Y	1.650	1.661	$r<v$
$Z\wedge c=5°–6°$	1.658	1.668	

Hebron, Maine. The first description from an American locality of the childrenite-eosphorite series was given by Brush (1863) for the material from Hebron. Because its physical, pyrognostic and crystallographic properties were similar to the earlier described childrenite, it was called childrenite. Cooke (1863) described the crystallography and oriented the crystals with the *b* axis vertical. Larsen (1921) determined the optical properties and, because of the relatively low indices of refraction, concluded the mineral was eosphorite.

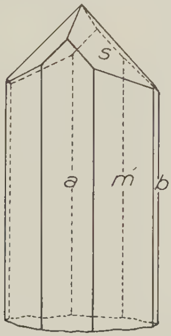

Fig. 5. Eosphorite. Buckfield, Maine.

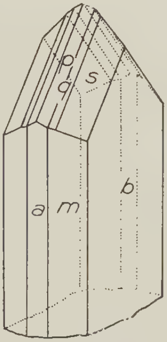

Fig. 6. Eosphorite. Hebron, Maine.

The writer obtained from the National Museum the same specimen from Hebron (U.S. National Museum specimen #82429) used by Larsen. The crystals are uniform in size, about 2 mm. wide and 4 mm. long. They are flattened on {010} with other forms *a*{100}, *m*{110}, *s*{121}, *q*{232} (Fig. 6). Although Brush and Dana (1878) show a figure of a doubly terminated crystal from Hebron, all those seen were attached at one end. The crystals are partially altered and manganese oxide coatings along cracks give them a black color and render them nearly opaque. The crystals are probably twinned, but because it is impossible to transmit light through a complete crystal, no twinning was observed. The alteration also made it impossible to tell whether there is zoning. However, the variation within a single crystal of the indices of refraction, as given below, indicates zoning. Only the tips of the crystals give the higher indices.

Optical Properties of Eosphorite, Hebron, Maine

Orientation	n(Na)	
X=b	1.631–1.635	Opt.—
Y	1.657–1.665	2V=∼50°
Z	1.663–1.671	$r<v$

A chemical determination gives 4.52% FeO. The specific gravity is 3.08 but may be low because of the altered nature of the crystals.

Poland, Maine. Drugman (1914) gives two analyses of eosphorite from Poland. An average of the two gives 5.15% FeO, 26.03% MnO. By the method of minimum deviation he determined $n\text{X}=1.642$, $n\text{Z}=1.669$.

A few isolated and broken crystals of eosphorite on specimens of other iron-manganese phosphates from Poland, Maine are all that are at present available in the Harvard collection. An optical study of these gave the following results: $n\text{X}=1.638$, $n\text{Y}=1.657$, $n\text{Z}=1.665$; $\text{X}=b$, $\text{Z}\wedge c=3°$, $2\text{V}=45°$, $r<v$. It is difficult to reconcile these refractive indices with Drugman's, for his method of determination should have been at least as accurate as that used by the writer. The specific gravity is 3.09.

Branchville, Conn. In their description of the minerals from Branchville, Brush and Dana (1878) give a crystal drawing of eosphorite. It is of a singly terminated crystal with the forms: $a\{100\}$, $b\{010\}$, $m\{110\}$, $n\{120\}$, $p\{111\}$, $s\{121\}$, $q\{232\}$. No crystals were available in the present study, for the only Branchville eosphorite in the Harvard collection is a rough radial aggregate. Consequently, the presence of twinning could not be determined.

Two chemical analyses by Penfield and one by Wells (Brush and Dana, 1878) give an average of 7.22% FeO and 23.15% MnO. Constant refractive indices indicate that the material is uniform and the optical properties determined by Larsen (1921) were verified. However, the specific gravity of 3.09, obtained both by suspension in heavy liquid and with the Berman balance, is at variance with G=3.134 given by Brush and Dana.

Black Mountain, Rumford, Maine. Eosphorite from Black Mountain found in cavities in feldspar is in excellent prismatic crystals ranging in length from 0.10 to 20.0 mm. The larger crystals are a reddish brown, the smaller are pale yellow. Thus in color there is a close resemblance to the crystals from Red Hill three miles away. The habit of the crystals from the two localities is somewhat similar, with the Black Mountain crystals showing the forms *a*{100}, *b*{010}, *m*{110} and *s*{121} (Fig. 7). In addition to the well crystallized material, a pale pink to colorless eosphorite also occurs at Black Mountain in coarse radial aggregates.

The indices of refraction of the crystals show a considerable range from base to tip indicating a compositional variation along the *c* axis. Several long crystals were broken and the specific gravity taken of the ends. For the base G = 3.104; for the tip G = 3.121. These values probably do not represent the true densityof either extreme end, for the pieces used were not of uniform composition as shown by the refractive indices. For the coarse radial eosphorite G = 3.10. Optically it appears to be very uniform, and a chemical analysis of it is given beyond.

Optical Properties of Eosphorite, Black Mountain, Maine

Orientation	Crystals Base *n*(Na)	 Tip *n*(Na)	Coarse Aggregate *n*(Na)	
X = *b*	1.638	1.642	1.640	Opt. (−)
Y	1.660	1.674	1.660	2V = 50°
$Z \wedge c = 6°$	1.667	1.680	1.667	$r < v$

Analysis 1 below is of the radial eosphorite from Black Mountain. For comparison are given analyses of eosphorite from Buckfield, Maine (2) and of childrenite from St. Austell, Cornwall, England (3).

Chemical Analyses of Eosphorite and Childrenite

	1	2	3
P_2O_5	30.91	29.89	29.92
SiO_2	—	.90	—
Al_2O_3	23.90	22.37	21.43
FeO	11.37	1.38	28.56
MnO	18.41	29.94	3.11
CaO	—	—	0.65
Na_2O	—	—	0.37
H_2O	15.26	15.34	15.80
F	—	trace	—
Total	99.85	99.82	99.84

1. Black Mountain, Rumford, Maine. F. A. Gonyer, analyst.
2. Buckfield, Maine. H. Vassar, analyst (Landes, 1925).
3. St. Austell, Cornwall, England. (H. Otto, 1935).

North Groton, New Hampshire. In the spring of 1949 eosphorite was discovered at the Palermo mine, North Groton, associated with other iron-manganese phosphates. The best specimen from this locality was found by Mr. Charles Weeks and was loaned to the writer for study. The crystals are transparent, of a light brown color and very similar in habit to those from Newry (Fig. 2), flattened on {010}.

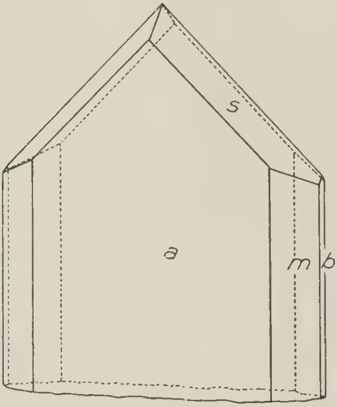

Fig. 7. Eosphorite. Black Mountain, Rumford, Maine.

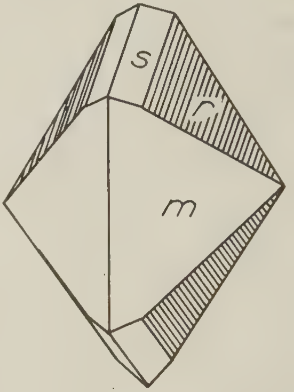

Fig. 8. Childrenite. Tavistock, Devonshire, England.

The Palermo eosphorite crystals show extreme compositional variation from the base to the tip as indicated by the refractive indices. All the crystals show twinning by reflection on {100} with an extinction angle angle against *c* of 8°. It was impossible to segregate separate portions of the crystals for iron determinations or for specific gravity measurements. For the crystals as a whole there is 12.15% FeO and G=3.12.

Optical Properties of Eosphorite, North Groton, New Hampshire

Orientation	Base of crystals n(Na)	Tip of crystals n(Na)		
$X=b$	1.641	1.653	colorless	Opt. −
Y	1.665	1.673	colorless	
$Z\wedge c=8°$	1.669	1.683	brown	$r<v$
	$2V=35°$	$2V=50°$		

Red Hill, Rumford, Maine. The Red Hill eosphorite is in cavities in feldspar associated with amythestine quartz. It is reddish brown in color. There are two generations of crystals; the earlier are the larger with a maximum length of about 1 cm. The later crystals which vary in length from 0.1 to 1.0 mm. are arranged in radial groups and as subparallel overgrowths on the earlier crystals. The habit of both generations of crystals is the same and similar to crystals from Black Mountain (Fig. 7) with the forms *a*{100}, *m*{110}, and *s*{121} and on some crystals *b*{010} poorly developed.

The indices of refraction, as given below, indicate that the earlier deposition, as in the zoned crystals at Newry and elsewhere, is lower in iron. Moreover, the specific gravity, which for the earlier crystals is 3.14 and for the later crystals 3.15, indicates a like compositional difference. In the larger crystals there is 14.62% FeO and 13.86% MnO. Because of insufficient material chemical determinations were not made on the smaller crystals but the curves in Fig. 11 would indicate about 20% FeO.

OPTICAL PROPERTIES OF EOSPHORITE FROM RED HILL, RUMFORD, MAINE

Early large crystals		Late small crystals	
Orientation	n(Na)	Orientation	n(Na)
$X=b$	1.640 colorless	$X=b$	1.644 colorless
Y	1.664 yellow	Y	1.673 pale yellow
$Z\wedge c=4°$	1.670 light brown	$Z\wedge c=2°$	1.679 yellow

For both crystals: Opt. −, 2V=45°, $r<v$, strong.

CHILDRENITE

Of the specimens available for study only those from England have a sufficiently high excess of FeO over MnO to be called childrenite. The average of four childrenite analyses by Drugman (1915) from the Crinnis mine near St. Austell, Cornwall gives 28.30% FeO, 3.35% MnO. Otto (1935) in an analysis of material from the same locality gives 28.56% FeO, 3.11% MnO. The other major occurrence is in the George and Charlotte mine near Tavistock, Devonshire. Three analyses of childrenite from Tavistock are available; Rammelsberg (1852), Church (1873), Penfield (1880). No indices of refraction are given for the first two, but Otto (1936) gives indices which he implies are for the childrenite analyzed by Penfield. The writer checked these indices and on the same material Gonyer determined 26.40% FeO. Penfield's analysis gives 26.54% FeO and 4.87% MnO.

Crystals from the two English localities, unlike crystals of eosphorite, are doubly terminated. Figure 8 illustrated a common habit at Tavistock and Fig. 9 at St. Austell. It was hoped that these doubly terminated crystals would show some morphological evidence of twinning that is lacking in the singly terminated eosphorite. However, as illustrated in the figures, the crystals appear to have perfect orthorhombic symmetry.

When a crystal is cut or broken so that it will lie parallel to {010}, it shows twinning under polarized light. Moreover, in addition to twinning by reflection on {100} as in eosphorite, it is also twinned by reflection on {001}. Between crossed nicols a {010} plate shows a fourling (Fig. 10) made up of two individuals.

It is not surprising that this twinning has not been noted previously since the {010} form is lacking or is poorly developed and it is only with difficulty that a crystal can be oriented in such a way as to permit the twinning to be observed.

The optical properties of childrenite can be summarized as follows:

Optical Properties of Childrenite

Orientation	Tavistock n(Na)	St. Austell n(Na)	
X=b	1.643	1.643–1.646	Opt. −
Y$\wedge c$=8°	1.678	1.676–1.681	2V=50°
Z	1.684	1.681–1.687	$r>v$, strong

Childrenite differs from eosphorite not only in having higher indices of refraction but also in dispersion and in optical orientation. In all eosphorite the dispersion is $r<v$; in childrenite $r>v$, strong. In eosphorite Z makes a small extinction angle with c; in childrenite Y makes the corresponding extinction angle with c. Although most of the crystals examined from St. Austell gave an 8° extinction angle, some portions of zoned crystals show an extinction angle as low as 1°. This low angle is on that portion with the lower refractive indices.

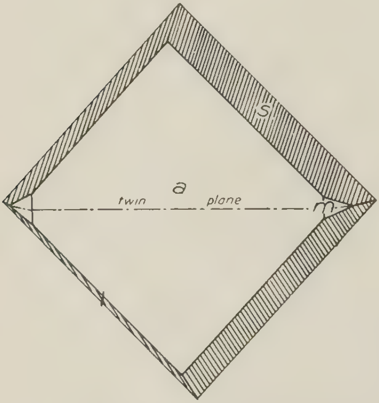

Fig. 9. Childrenite. St. Austell, Cornwall, England.

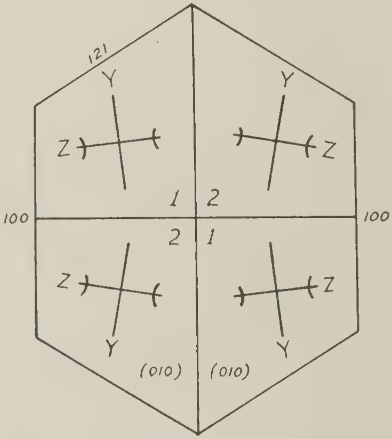

Fig. 10. Childrenite lying on (010) showing optical orientation in fourling.

The specific gravity of childrenite has been given as high as 3.24. However, several determinations gave consistant results as follows: Tavistock G=3.18, St. Austell G=3.19. The calculated value of Barnes (1949) for childrenite is G=3.186.

Other Occurrences. Gordon (1944) describes childrenite from Llallagua, Bolivia, and gives a chemical analysis showing 21.69% FeO and 4.21% MnO. However, his indices of refraction, $n\text{X}=1.652$, $n\text{Y}=1.660$, $n\text{Z}=1.678$, give the mineral a *positive* optical sign which is at variance with the sign as determined on all other members of the series. His specific gravity of 3.05 is also low for such a high iron member.

Mason (1942) reported the presence of eosphorite in a specimen from Hühnerkobel, Bavaria. Since he gives no chemical analysis and the material was not available for an iron determination, his data are not listed in Table 2. The indices of refraction, $n\text{X}=1.642$, $n\text{Y}=1.663$, $n\text{Z}=1.669$, fit the curves in Fig. 11 reasonably well and correspond to an eosphorite with about 12% FeO. The specific gravity is given as 3.07.

Table 2. Summary of Optical Data
Eosphorite-Childrenite Series

Locality	% FeO	% MnO	nX	nY	nZ	2V	Orientation $\text{X}=b$	Dispersion	G
1. Buckfield, Maine	1.38	29.94	1.629	1.650	1.658	45°	$\text{Z}\wedge c=5^\circ$	$r<v$	3.10
2. Hebron. Maine	4.52		1.631	1.657	1.663	50°		$r<v$	3.08
3. Poland, Maine	5.15	26.03	1.638	1.657	1.665	45°	$\text{Z}=c=3^\circ$	$r<v$	3.09
4. Branchville, Conn.	7.22	23.15	1.633	1.656	1.664	51°		$r<v$	3.09
5. Newry, Maine	10.10		1.638	1.660	1.667	50°	$\text{Z}=c=6^\circ$	$r<v$	3.10
6. Black Mt., Rumford, Maine	10.26		1.640	1.660	1.667	50°	$\text{Z}\wedge c=6^\circ$	$r<v$	3.11
7. North Groton, N. H.	12.15		1.641	1.665	1.669	35°	$\text{Z}\wedge c=8^\circ$	$r<v$	3.12
8. Red Hill, Rumford, Maine	14.62	13.86	1.640	1.664	1.670	45°	$\text{Z}\wedge c=4^\circ$	$r<v$	3.14
9. Tavistock, Devonshire, England	26.54	4.87	1.643	1.678	1.684	50°	$\text{Y}\wedge c=6^\circ$	$r>v$	3.18
10. St. Austell, Cornwall, England	28.56	3.11	1.646	1.681	1.687	50°	$\text{Y}\wedge c=8^\circ$	$r>v$	3.19

The curves of Fig. 11 made from the data summarized in Table 2 show a rather uniform increase in refractive index with increase in FeO. The birefringence also increases slightly with FeO. No analysis, with the exception of Gordon's, shows percentage of FeO between 14.62% and 26.56%. For this reason one might suspect that the series is one of limited isomorphism. However, the refractive indices of the zoned eosphorite crystals argue against this. If the higher sets of indices of these zoned crystals were plotted on the curves, they would fall within the gap, indicating that a portion of certain crystals have this intermediate composition. Both the high and low index parts of the zoned crystals have the same optical orientation and dispersion. Only in the high iron childrenite from England is there a change in the optical orientation and a reversal in dispersion.

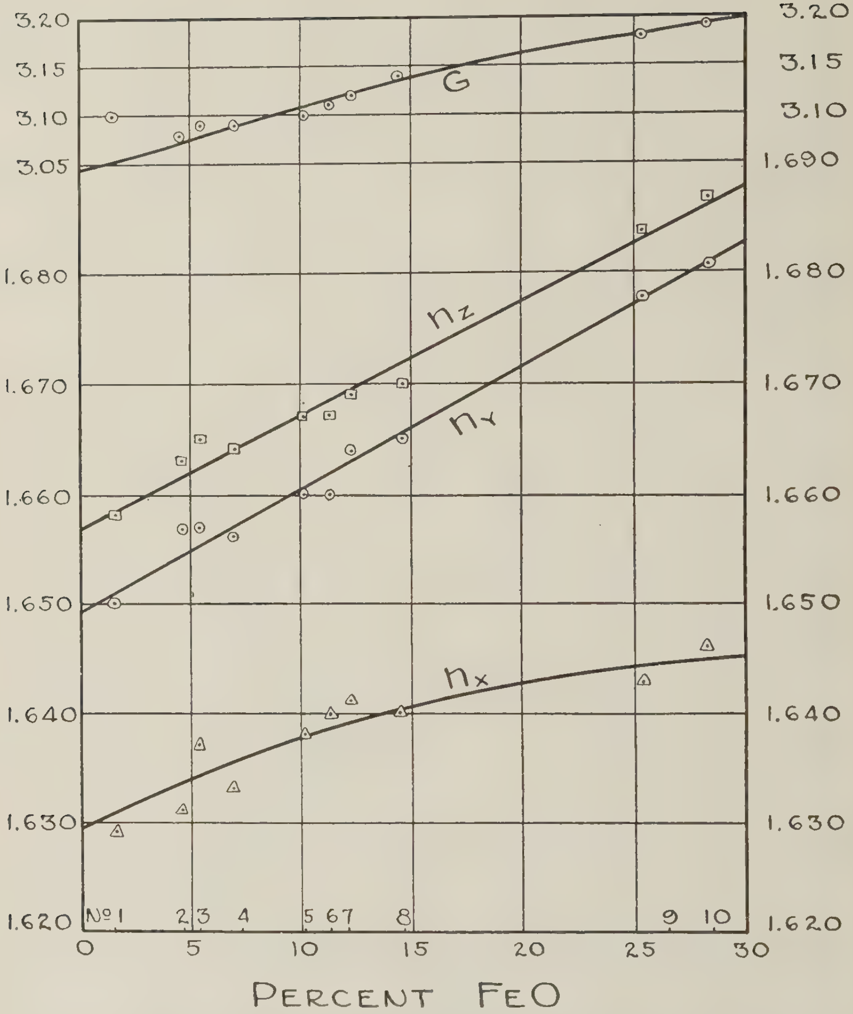

FIG. 11. Variation of indices of refraction and specific gravity with composition in childrenite-eosphorite series.

With only one marked exception, the specific gravity shows a regular increase with increase of FeO. Extrapolating from the curve, for the pure iron end-member G = 3.20, for the pure manganese end-member G = 3.05.

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THE WATER CONTENT OF PRIMITIVE GRANITIC MAGMA*

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Abstract

Previous estimates of water content of magmas are summarized. Assuming that essentially the amount of water now on the Earth's surface was in a primitive atmosphere and that this atmosphere was in equilibrium with a molten outer layer of the Earth as it began to crystallize, the amount of water in this "primitive magma" can be calculated. It comes out about 3%.

Introduction

An estimate of the amount of water in magma depends directly upon certain assumptions, such as the mechanism of derivation of the magma, physical conditions in the surrounding medium, relation of the magma chamber to the surface, and conditions at the surface. It is also likely to depend indirectly upon an individual's idea of how much water must have come from a given rock mass to produce certain observed effects; in other words, on his idea of how much water the magma *should* have contained. For these reasons estimates of water in magma have ranged from almost nil to many weight per cent.

Previous Estimates of Water in Magma

Green (1887) and later Brun (1911) maintained that magma, at least where it comes to the surface in volcanoes, is essentially anhydrous. Most other workers have concluded that water is usually, if not always, the chief agent in volcanic activity and that it is also important in the activity of intrusive magmas that do not reach the surface.

In 1847 Scheerer (p. 491) estimated that the water content of granitic magmas was between 1 and 50% and suggested that the true value was probably nearer the minimum figure—possibly 5, 10, or 20%.

Lindgren (1905) concluded from the fact that pitchstones contain up to 8% of water that the magmas from which they were derived probably contained a like amount but that "all magmas may, of course, not have contained this amount." Allen (1922) used the same argument in discussing a pitchstone containing 10.5% of water, . . . "Since hygroscopic water in rocks of this nature is insignificant, and since the rock showed no sign of alteration, we may infer that most of this water was contained in the original magma." Hawkes and Harwood (1932), however, in discussing the glassy border phase of a dike that contains 5.88% of water, arrived at a quite different conclusion. They had evidence that the dike could not have been more than 0.5 km below the surface at the time of intrusion and pointed out that, from Goranson's data, (1931), the magma could not have contained more than 1 to 2% of water at the pressure prevailing at that depth and at a temperature of 900° to 700° C. They concluded, therefore, that "the bulk of the water now in the glass was not held in the magma on emplacement, but has been added later." They find it "reasonable to suppose that the water was added from the interior of the dike, from which it was expelled on crystallization."

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Vogt (1922) arrived at a figure for water in the "initial parent magmas" by assuming that the water in the oceans plus that in the crust (2.02%) was originally in solution in a shell 30 miles thick. On these assumptions the original water content would have been 4.06%. Vogt believed, however, that some water would have escaped before crystallization began and that the water content of the derived "partial" magmas was always less than 4% when crystallization began.

Tyrrell (1926, p. 47) estimated that the basaltic magma of the Hawaiian Islands contains about 4% of water.

Grout (1928) suggested that 4% might be a valid figure for the water content of the magma from which the diabase sill on Pigeon Point, Minnesota, was formed. There is no direct evidence on which to base such an estimate, but the olivine diabase contains well over 1% of water and "many observations show conclusively that magmas contain more water than the igneous rocks formed by their crystallization" (p. 568). Grout cited the 4% estimate of Tyrrell and concluded that the estimate of 4% of water in the primary magma of the sill was therefore not unreasonable.

In this same paper (p. 568) Grout says, "Acid rock magmas, such as form pitchstone and pegmatitic granite, may contain 10 to 20% of water." He does not support the statement with examples, literature references, or a theoretical discussion.

Bowen believes (1928, p. 301) that the average basalt magma may not contain more than 0.5% of total volatiles originally, which would, of course, mean a water content of less than 0.5%. He suggested (p. 302) that with crystallization of anhydrous phases and consequent accumulation of water, the water content may rise to a maximum of 4 or 5%. For magma with the composition of granite the estimate is less than 1% of volatiles (p. 296). He indicated (p. 297) that exceptional magmas (such as gave rise to pitchstones) may have contained water up to the order of 10%. He believes that such cases are decidedly abnormal and that the normal granitic magma contains much less water, probably less than 1%.

Later (1933, p. 117) Bowen used a hypothetical example in which he assumed 0.5% original water content of a granitic magma. He thinks that after the water content has reached 2 or 3%, hydrous phases may form by reaction so that the water content never increases beyond this value, and the final product would be a granite with 0.5% of water.

Goranson (1931) determined experimentally the amount of water that would dissolve in molten Stone Mountain granite under various pressures at 900° C., which is a reasonable figure for the beginning of crystallization in a granite. The amount of water that dissolved in the melt approached 10% as the pressure was increased, being 9.35% at 4000 bars. The curve flattened rapidly above 3000 bars. In applying these results to magmas he assumed that crystallization took place from the center of the Earth outward, progressively expelling the water until most of it was concentrated in a molten granitic layer some 15 km thick adjacent to the surface. *If* there was enough water present to saturate this layer, the amount of water at a given spot depended upon the composition of the (anhydrous) melt, the temperature, and the pressure.

From this postulated setup and his experimental work, Goranson calculated the water content at various depths, assuming granitic composition and a temperature for each depth considered. He concluded that if these conditions represent rather closely those obtaining in nature, rhyolite glasses could not very well have more than 10% of water and that glasses with 8 to 10% of water should be extremely rare. He pointed out that this conclusion appeared to be supported by analyses of fresh obsidians.

Phemister (1934) estimated that plutonic basaltic magmas contain 4 to 5% of water and that similar magmas that are less deep-seated contain less water in solution. This figure is obtained indirectly. Phemister made a detailed study of crystallization of a hydrous melt in the system K_2SiO_3-SiO_2-H_2O (from Morey's diagrams) to evaluate the external factors affecting crystallization of an "ideal" basalt magma and arrived at the above-mentioned estimate for the amount of water in such a magma.

Gilluly (1937) estimated that basaltic magmas may contain about 4% of water; granitic, about 8%. These figures are based on evaluation of geologic data. Gilluly believes that water contents of this order of magnitude are necessary to explain observed activity of volcanoes, and the amounts of migmatization and granitization that have accompanied many granitic intrusions.

Hess (1938) suggested that some ultrabasic magmas may contain as much as 5 to 15% of water in solution. This is based on his idea that serpentinization takes place during the later part of the crystallization of the magma. A completely serpentinized rock contains about 13% of water, which is the same order of magnitude as Hess' maximum figure. If Hess' ideas of the origin of such rocks is correct, then magmas with such water contents must have existed.

Grout (1941) took 10% as the part of a magma that may have been volatile. For simplicity he assumed that this is water carrying 10% of mineral matter in solution. Apparently the 10% water was taken because it accords roughly with the amount (8%) estimated earlier by Gilluly (1937). Assuming 10% of mineral matter in solution gives 1% as the amount of the nonvolatile portion of a granitic intrusion that might be mobile in this manner, Grout pointed out that this is the proportion of pegmatite in granite estimated by Vogt (1931).

Daly (1944) extrapolated Goranson's curves to 23,000 atm (minimum pressure on the basaltic substratum) and decided that at this pressure granitic magma could contain 12% of water. He believes that under the same conditions basaltic magma might contain 5 to 10% of water. In the absence of experimental work on materials of basaltic compositions, Daly assumed that basaltic magma would dissolve about half as much as granite, i.e., magma in the basaltic substratum would carry 6% of dissolved water.

Daly pointed out that as the magma ascends into the Earth's crust water is likely to escape. He concluded, apparently in part from his estimate of original water content and in part on the basis of analyses of fresh basalt flows, that basaltic magma may reach the Earth's surface with as much as 2 or 3% of water and may retain this amount even after solidifying.

Graton (1945) cited earlier estimates of water content of magmas by Bowen, Gilluly, and Goranson, and concluded that the solubility *limits* in nature probably range from 5 to 11% of volatile content for the various common magmas, but that actual water contents of lavas as brought to the surface in volcanoes are consideraly less. He suggested that when the water content of lava is not more than 2 or 3% quiet extrusions take place, but that when the water content exceeds a critical value of about 3% explosive ejection results. Goranson's results undoubtedly had some influence on Graton's choice of 11% as the upper limit, but another consideration also entered. He calculated that the expansion and escape of still larger proportions of volatile constituents would produce such a cooling effect that the magma would be chilled almost immediately and bring activity to an end. Since this effect is not observed in volcanoes, the magma cannot have contained much more than about 10% of water.

BASES OF PREVIOUS ESTIMATES

All of these estimates of the water content of magma can be divided into a few groups according to the basis on which the estimates were made:

(1) On the basis of the water content of igneous rocks.

(2) On the basis of the observed effects of emanations from igneous rocks.
(3) Experimental determination of solubility of water in silicate melts.
(4) Assuming that all water now at the surface has been derived from crystallizing magma. A thickness for the shell from which the water was derived must be assumed, of course, in order to obtain an estimate for original water content.
(5) Combinations of (1) to (4).

ASSUMPTIONS FOR PRESENT CALCULATION

It is interesting to start with an entirely different set of assumptions, as nearly as possible independent of the considerations mentioned above, and to see how the results compare with previous estimates. Let us assume:

(1) That, at an early period in the Earth's history the outer layer, at least, was molten.
(2) That essentially the same amount of water was present on the Earth's surface then as we find today in the hydrosphere, the biosphere, the atmosphere, and the sedimentary rocks.
(3) That, at the time crystallization began, equilibrium had been established between this primitive atmosphere and the molten crust —the original magma from which at least most of the igneous rocks found at the surface today have been derived either directly or indirectly.

The first assumption has purposely been simplified. This simplification is probably justified. There is ample geophysical evidence of a compositional-density zoning in the Earth, so if convection has been operative the currents probably have been confined to individual layers and would not have extended from the surface into subcrustal layers.

There is no way to make an accurate check on the second assumption, but the evidence is that the ocean basins have had about their present area and volume throughout geologic history and that they have been filled to about the same level with water during all that time. The earliest known pre-Cambrian sedimentary rocks are widely spread over the continental masses. If the amount of water on the Earth's surface had been increasing or decreasing steadily and significantly throughout geologic time then there should have been a general and progressive increase or decrease in the area covered by progressively younger sedimentary rocks. Because of coverage and erosion of large amounts of sedimentary rocks, it is not possible to estimate areas covered by rocks of various ages very accurately, but the paleogeographic maps thus far compiled fail to show any such consistent trend in either direction.

As to the third assumption, under the postulated conditions equilibrium would obtain at the surface of contact between the molten crust and the primitive aqueous atmosphere. The (assumed) thickness of the molten layer saturated with water under these conditions in no wise affects the calculations that follow, because the water on the surface is that which was left *after* equilibrium was established. Increasing pressure with increasing depth has no meaning in this situation, as it has for Goranson's and for Daly's calculations, because the water comes from the outside and it is not possible to get more water in the lower part of the layer than the equilibrium amount at the surface. Even if the vapor pressure of the water could act directly on the lower parts (which it could not through a viscous glassy layer) the pressure of the magma would have no effect. The difference would be due to the increased height of the water-vapor column and would not be significant.

Calculation of Water Content

Granted the assumptions, the calculations are straightforward and give a value that is subject to relatively slight uncertainty.

At the temperature where crystallization would begin, all the water on the surface would be in a gaseous envelope. The pressure per unit area of the superincumbent water vapor would be the total weight of the water in the gaseous envelope divided by the area of the interface, i.e., the Earth's surface. Rubey (unpublished manuscript) estimates that all the water in the hydrosphere, atmosphere, biosphere, and sedimentary rocks amounts to 1.67×10^{24} g. The Earth's surface is approximately 5.1×10^{18} cm², so the pressure of the water vapor per unit area would have been

$$\frac{1.67\times10^{24}}{5.1\times10^{18}}=3.3\times10^{5}\ \mathrm{g/cm^2}=330\ \mathrm{kg/cm^2}.$$

Anhydrous granite melts at about 1200° C. Under 1000 bars pressure of water vapor it melts at about 700° C (Goranson, 1931, p. 499). If we use these two figures and draw a curve shaped like the one for melting-point lowering of sodium disilicate with increasing water-vapor pressure (Fig. 1), it will be seen that under a pressure of 330 kg/cm² granite would begin to crystallize in the neighborhood of 900° C. According to Goranson's experimental results (1931, Fig. 3) a granite melt under these conditions would contain very close to 3% of water.

O. F. Tuttle, of the Geophysical Laboratory, very kindly made two runs of Stone Mountain granite with water at 725° C and 2000 kg/cm². In both runs the product was entirely glassy, containing 6.1 and 6.7% H_2O respectively. The technique was entirely different from Goranson's, but the solubility indicated is of the same order of magnitude. Tuttle's lower results may be explained by the short duration of the runs—2 hours—because water penetrates viscous high-silica melts very slowly. Charles E. Kinser, of the U. S. Geological Survey, analyzed the samples for water.

Under the above assumptions, therefor, the primitive granitic magma underlying the continents would have contained approximately 3% of water. No experimental results are yet available for solubility of water in basaltic or peridotitic melts. If and when such experiments are performed the water content of primitive magmas of these compositions can be derived from the above assumptions by analogous calculations.

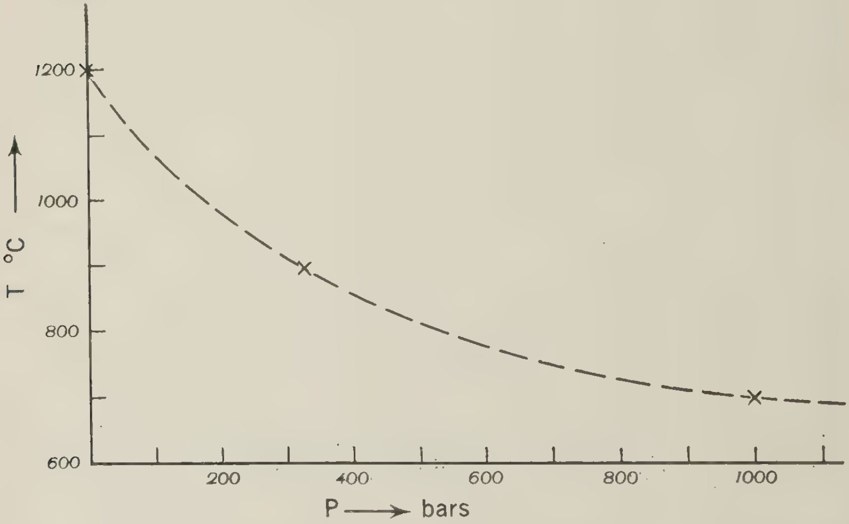

FIG. 1. Curve showing the decrease in melting point of granite with increasing pressure of water vapor. The points at atmospheric pressure and at 1000 bars were determined experimentally by Goranson (1931). The shape of the curve is patterned after that of the system $Na_2Si_2O_5$-H_2O. At 300 kg/cm² the melting point of granite would be very close to 900° C.

After crystallization had proceeded to a point where a solid shell existed and at least a considerable part of the water from the primitive atmosphere had condensed the problem would be quite different. It is probable that the *average* magma would still contain approximately the original 3% of water, but locally derived secondary magmas might vary considerably in either direction.

For example, a body of magma near the surface at low pressure might lose quite a lot of its water and so have less than 3%. Or there might be local intrusions of relatively anhydrous magma squeezed up from the underlying layer that was never in equilibrium with the aqueous atmosphere.

On the other hand, crystallization of anhydrous phases might concentrate water in a rest magma that could be filter-pressed away and so produce a secondary magma with much more than 3% of water. Fusion of depressed hydrous sediments might produce a similar result.

The question of the water content of secondary magmas is in each case a separate problem and all available data must be sought and used in its solution.

SUMMARY AND CONCLUSION

TABULAR SUMMARY OF ESTIMATES OF WATER IN MAGMA

Date	Author	Type of magma	% Water	Method of estimating; assumptions; remarks
1847	Scheerer	Granitic	1 to 50 5, 10, or 20 probable	
1887	Green		0	
1905	Lindgren	High silica	Up to 8	Water content of pitchstones.
1911	Brun		0	
1922	Allen	High silica	Up to 10	Water content of pitchstones.
1922	Vogt	Initial parent	4	Water in and on crust originally uniformly distributed in 50-km shell.
1926	Tyrrell	Basaltic	4	
1928	Grout	"Diabasic"	4	Present water content of diabase sill+Tyrrell's estimate.
1928	Grout	"Acid" and pegmatic	10 to 20	
1928	Bowen	Basaltic	<0.5	Original.
			4 to 5	Maximum after crystallization of anhydrous minerals.
1928	Bowen	Granitic	<1.0	Common case; most obsidians <1.0% water.
			8±	Exceptional; some pitchstones show such quantities of water.
1931	Goranson	Granitic	<10	Exp. det. of solubility of water in granite melts; calculation for estimated temperature and pressure at various depths in Earth.
1933	Bowen	Granitic	0.5 orig. 2 to3 max.	Hypothetical example.

Tabular Summary of Estimates of Water in Magma (*continued*)

Date	Author	Type of magma	% Water	Method of estimating; assumptions; remarks
1934	Phemister	Basaltic	4 to 5	Analogy with system K_2SiO_3-SiO_2-H_2O.
1937	Gilluly	Basalt	4	Geologic data, volcanoes, etc.
1937	Gilluly	Granitic	8	Geologic data, granitization, etc.
1938	Hess	Ultrabasic	Up to 15	Field evidence on serpentinization and related phenomena.
1941	Grout	Granitic	Up to 10	Accords with Gilluly; can explain 1% pegmatite estimated by Vogt.
1944	Daly	Granitic	12	Extrapolates Goranson's results to 23,000 atm (base of crust).
1944	Daly	Basaltic	6	Assumes that one-half as much water will dissolve in basaltic melt as in granitic.
1945	Graton	"Common"	5 to 11 volatiles	Goranson's results plus consideration of cooling effects of escape of larger quantities of volatiles.
			3	Approx. limit for nonexplosive volcanic activity.
1950	Ingerson	Granitic	3	Assumes equilibrium between primitive atmosphere containing all water at present on Earth's surface and a molten outer layer.

It is concluded that, if the assumptions made are valid, the primitive (granitic) magma of the Earth's outer layer contained approximately 3% of water. Later magmas derived directly or indirectly from the primitive one or produced in other ways may have contained either more or less water. It is probable, however, that any marked deviation from the original 3% is unusual and that most magmas contained approximately this amount of water when they began to crystallize.

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RADIOACTIVITY AND MINERAL DEPOSITS

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ABSTRACT

Radioactivity is important in a variety of problems in economic geology. Radiations from radioactive ore deposits provide a means of locating shallow ore-bodies. Anomalous distribution of radioactivity also may be of limited use in locating faults. Variations in radioactivity of ordinary rocks are proving useful as a means of correlating rocks, and of logging cased wells, and radioactivity produced in the more common rock-forming elements by artificial neutron bombardment likewise is proving useful as a logging method.

Studies of rocks show an increase in the proportion of radioactive elements in differentiation from basic rocks to granites and pegmatites, and have provided some insight into the mechanism of formation of radioactive ore. In non-radioactive ore-bodies, the small variations in radioactivity appear to be of no genetic significance, but occasionally may prove useful for purposes of correlation.

The rates of accumulation of helium and lead calculated from radioactivity are useful in conjunction with helium, lead, or isotopic analyses for the calculation of geological age. The helium method has the weakness that helium escapes from altered and open-structured minerals during geological time, but enough is now known about the limitations of the method to permit its use on selected material, thus reducing the error enough to be useful in problems where the age is not known by geological criteria within wide limits. The lead method, now restricted to rare radioactive minerals, may eventually be developed to the point where it may be applied to many granitic and other type rocks.

Studies of the radiogenic and isotopic history of lead ores suggest that hydrothermal ores of this type are derived from rocks or rock magmas.

GEOLOGICAL AGE

Helium Method:

Calculation of geological age from the concentrations of radioactive elements and accumulated helium (Fig. 1) is complicated by the variable degree of retention of helium by different rocks and minerals. In an unaltered Devonian granite of expected age about 300 million years, one might find: granite—helium index 90; quartz—75; feldspar—60; biotite—180; hornblende—250.

Such figures, obtained from substitution of helium and radioactivity analyses in the age equation, are clearly not absolute ages, but it is useful to express the results in this manner, using the term *helium index.*

If all granitic rocks showed nearly the same degree of retention of helium, determinations of the helium index would be useful in geological correlation, even though absolute ages were not established. As a matter of fact, rocks of one type with comparable geological histories have given the correct sequence in some cases, the best results having been obtained with basic crystalline rocks. Table 1 summarized a series of determinations on such rocks. Lane and Urry (1935) and Urry (1936, p. 1217) had previously found a measure of regularity in results from basic rocks.

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A statistical analysis of helium age results shows that there are only about two chances in three of assigning such a rock to the correct geological period. Not only is the problem complicated by the escape of some of the helium but occasionally excess helium is present as in a diabase from Yellowknife, N.W.T.: Yellowknife diabase—1000; feldspar—100; pyroxene—1650; olivine—800; magnetite concentrate—4800; expected helium age—600.

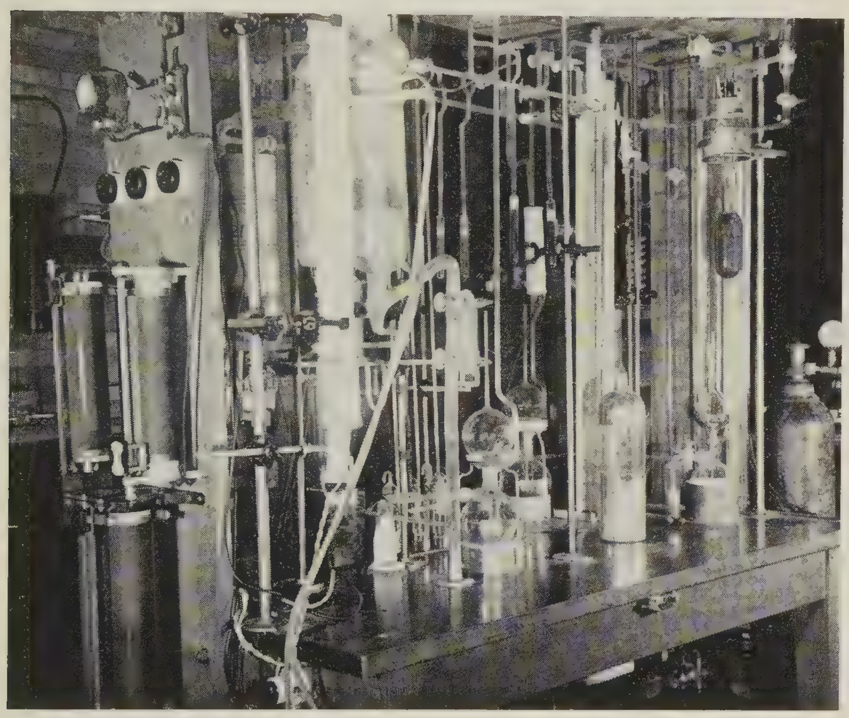

FIG. 1. Apparatus for the micro-analysis of helium.

Excess helium in rocks, however, seems to be rare, and it may be suspected by the absurdity of results, particularly when several minerals in a specimen have been studied. Geochemical considerations, together with a survey of several hundred helium indices, show excess helium most likely to be present in carbonates, and late hydrothermal minerals. In the case of the Yellowknife diabase the magnetite, associated with the titaniferous minerals, ilmenite and leucoxene is probably of deuteric origin.

In search of an answer to questions regarding retention of helium a number of minerals and rocks have been studied (Keevil, 1941) and it has been found that helium retentivity varies for different minerals, and that it may be affected by geological environment.

A theoretical treatment of the problem of diffusion of helium (Keevil, 1940) has suggested that it is desirable to use close-packed minerals with as nearly perfect structures as possible, with crystal dimensions greater than the range of alpha-rays, and with no excessive localization of radioactivity. Some experiments with such minerals have led to fairly promising results. As one example, Dr. H. H. Hess and the author have obtained the following results using carefully selected unaltered minerals: Augite—Gilboa Quarry, Triassic diabase, 182; bronzite—malips Drift bronzitite, 742; augite—Sekukuni norite, 588; hypersthene—Sekukuni norite, 511, suggesting that most of the radiogenic helium was retained during geological time, and Hurley and Goodman have found concordant results with a series of magnetite samples (Hurley and Goodman, 1943).

TABLE 1. HELIUM AGES OF SOME BASIC ROCKS

Location	*Helium Index*	*Geological Age*	*Helium Age*
Yellowstone Park	12	Miocene	21
Newfoundland	53	Triassic	105
Quebec	61	Triassic	120
New Jersey	68, 69	Triassic	140
New York	89	Triassic	175
Newfoundland	65	Pre-Triassic	150
Quebec	80	Pre-Triassic	185
Scotland	80, 109	Post-Devonian	220
New Jersey	150	Post-Ordovician	340
Newfoundland	321	Ordovician	530
Wyoming	200	Keweenawan	575
Michigan	186	Keweenawan	530
Quebec	210	Keweenawan	600
Wyoming	1550	Early Pre-Cambrian	—
Manitoba	1800	Early Pre-Cambrian	—

Loss of helium and the presence of excess helium has been found in all of these minerals, and no criterion for the selection of perfect material has been definitely established.

No two geological samples are identical, and the problem is further complicated by the high degree of localization of the radioactive elements in minor accessory minerals (Keevil et al., 1944) which may be present as sub-microscopic inclusions in otherwise perfect minerals. Nevertheless, the use of carefully selected unaltered minerals for helium age determinations shows some promise (Hurley, 1950).

Another method of attack is to correct helium indices for helium retentivity. Such a method has been followed in calculating the helium ages given in the last column of Table 1. Here, the helium indexes have been corrected on the basis of the average retentivity of basic rocks and its change with geological time. This method does not seem to be generally applicable, however, because of other variables which determine the final degree of retention. For example, an apparent relationship has been noted between degree of alteration, as determined with the petrographic microscope, and helium retentivity; hornblende unaltered—90% helium retained; ¼ altered—75% retained; ½ altered, 40%; ¾ altered, 15%; altered to chlorite—5% retained. A yardstick which measures the degree of alteration is needed for this application, but this does not seem to be forthcoming until more knowledge of the effect of metamorphism on storage of helium is at hand.

In either case, it is clear that the helium method cannot be 100% reliable, and aside from further development research, present application of the method to rocks would seem to be warranted only where the age difference in question is appreciable, or where geological criteria of age are lacking.

Lead Method:

More hope of obtaining absolute ages has been held for the lead method based upon the same radioactive mechanism, but using the more stable end-product lead as a measure of time. Up to the present, however, only highly radioactive minerals have yielded enough radiogenic lead for study. This has the dual weakness of limited application because such minerals are not always available in most problems of correlation, and well-dated material is too scarce for adequate means of standardization. Furthermore, selective leaching of uranium, thorium, or lead, frequently introduces difficulties. In spite of this, results by the lead method have been consistent enough to give confidence in the method.

The scope of the lead method has been increased by the discovery of extreme localization of radioactive elements in accessory minerals of some rocks. It is probable that this is quite general, and a few milligrams of radiogenic lead might be separated from concentrates of minor minerals and subjected to isotopic analysis by means of the mass spectrometer. This may make the lead method of age determination applicable to some rocks, particularly to the important group of acidic and intermediate intrusives.

A still more promising possibility, suggested by the above localization of radioactivity, and by the discovery of E. S. Larsen that lead is concentrated in weakly radioactive potassium minerals, is that the age may be determined directly from a spectrochemical analysis for lead and a determination of radioactivity of a concentrate of accessory minerals. This method would be based upon the premise that the correction for common lead in radioactive accessory minerals would be negligibly small. A joint research program to test this method is now under way (Keevil, Larsen and Harrison, 1949).

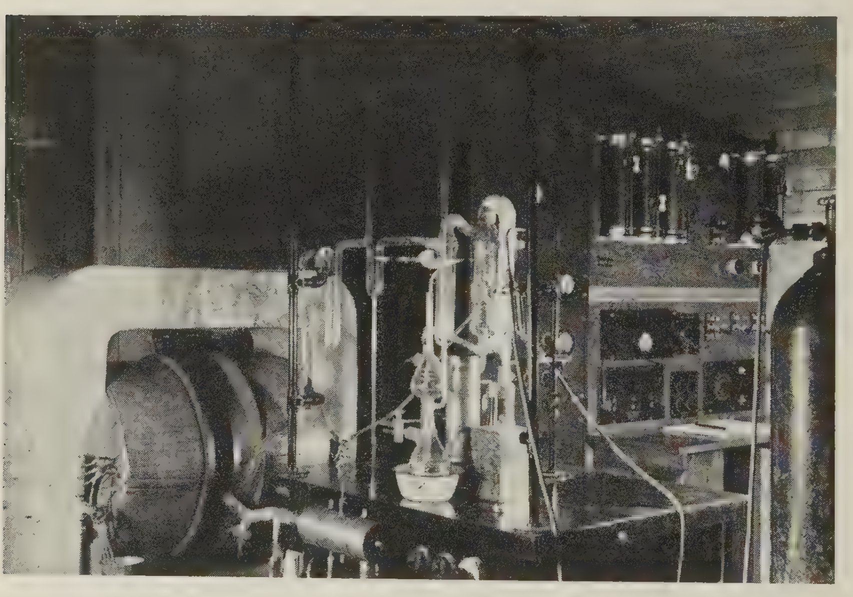

FIG. 2. Mass spectrometer for the isotopic analysis of lead.

DISTRIBUTION OF RADIOACTIVITY AROUND "NON-RADIOACTIVE" ORE DEPOSITS

General:

All of the districts included in this part of the investigation are usually considered to be of a non-radioactive type. The extreme sensitivity of radioactive experimental techniques makes possible the rapid quantitative determination of a few parts per million of radioactive elements. The results accrued are thus of geochemical as well as geophysical interest.

Porcupine Area, Ontario:

In the Porcupine area specimens from the Hollinger Dome, Preston East Dome and Coniaurum Mines were studied (Keevil, 1943). The quartz porphyries were found to be low in radioactivity, and no significant variation was observed in samples taken at various distances from ore deposits. The activities of 22 quartz porphyries, representing the Pearl Lake, Acme, Crown, Millerton, Miller Lake, Dome and Preston East Dome porphyries, as well as others in Bristol Twp., Tisdale Twp., averaged 0.42 and ranged in activity from 0.26 to 0.67 alpha-rays emitted per milligram per hour; similarly, 66 samples of porphry collected by E. L. Evans from the Hollinger Mine averaged 0.47. The darker-colored varieties tended to be more radioactive. Lower than average values were obtained for a specimen from the centre of the Coniaurum mass (0.28), and from the centre of the Pearl Lake mass (0.35), suggesting a distribution of radioactivity similar to that found by Ingham and Keevil (in press) for granodiorite and granitic batholiths (see also Wilson, 1949).

Two gold-quartz veins were only slightly more radioactive (0.10, 0.20) than a barren quartz vein (0.06), and two impure quartz samples separated from quartz porphyries (0.19 and 0.23) suggests a genetic relationship between gold-quartz and quartz porphyry. Similarly, the close similarity of the radioactivity of quartz porphyry pebbles in the conglomerate (0.40) with the average for 22 porphyries (0.42) indicates that the conglomerate is younger than the quartz-porphyry.

Newfoundland:

Typical ore and rock samples from the Buchans district, Newfoundland, were used for age determination (Keevil, 1943) and additional radioactivity measurements were made on other specimens. The activity of chalcopyrite in the Lucky Strike Mine (2.24) and in boulders (1.58) and mixed ore (1.40) is somewhat higher than usual, but a normal range of activities was found for other rock and mineral types in the area.

Franklin, New Jersey:

A study of the distribution of radioactivity in the Franklin and Sterling Hill area, New Jersey, and age determinations on a number of rocks and minerals have been made (Keevil, 1943).

Radioactivities of the rocks and pegmatites are summarized in Table 2. Aside from high values for the pyroxenites and basic pegmatites deep in the mines, the radioactivities are fairly characteristic of the rock types represented. The higher activity of blue limestone mixed with shale, and the lower activity when mixed with chert, are not unexpected since shales tend to be considerably more active than siliceous deposits; and the low activity of the pre-Cambrian gneiss is consistent with its antiquity. The only contact relationship noted was a tendency for the radioactivity of the white limestone to be higher near the pegmatites.

Table 2. Radioactivities of Rocks and Pegmatites at Franklin and Sterling Hill, New Jersey
alphas/mg./hr.

Rock	*Franklin, N. J.*	*Sterling Hill, N. J.*
Trap dike	0.2–2.2 (1)*	—
White limestone	1.53 (11)	0.57 (9)
Graphitic white limestone	1.1 (1)	—
Quartzite	—	0.7 (1)
Blue limestone	0.72 (3)	0.60 (4)
Blue limestone with shale	1.9 (1)	—
Blue limestone with chert	0.4 (1)	—
Pegmatites	—	1.0 (4)
Boni pegmatite	7.1 (1)	7.5 (1)†
Pyroxenites	—	0.7–44.0 (2)
Mica amphibolites	1.2 (4)	—
Pochuk gneiss	0.58 (6)	1.2 (1)

* Numbers in parentheses refer to number of specimens analyzed.
† Mafic portion 10.5; felsic portion 5.9.

Determinations of the radioactivity of ninety mineral specimens from the Franklin area are summarized in Table 3. The highest and lowest activities were obtained for minerals from the Franklin mine, and the average for Franklin minerals was somewhat higher than that for specimens from Sterling Hill, but on the whole the activities of minerals in the two mines were comparable both in magnitude and distribution.

The high radioactivity of certain minerals and the wide range of ac-

Table 3. Summary of Radioactivity Data on Minerals from Franklin and Sterling Hill, New Jersey

Minerals	*Activity*
	alphas/mg./hr.
Apatite, barylite and siderite with iron olivine (separated from magnetite)	7
Vesuvianite, beryllium-vesuvianite and apatite	3.2–4.6
Schefferite, magnetite, black roepperite and sphalerite	2.0–3.0
Zincite, sphalerite, manganiferous calcite, barysilite, calcium larsenite and calcium rhodonite	1.5–2.0
Franklinite, willemite, sphalerite, tephroite, zincite, hardystonite, galena	0.9–1.5
Franlinite, calcite, barite, garnet	0.8–1.2
Willemite, friedelite, franklinite, barysilite	0.4–0.8
Franklinite, willemite, zincite, manganiferous calcite, hancockite, barite, fowlerite, hardystonite, sphalerite, calcite, roepperite, cyprene, feldspar	0.2–0.4
Franklinite, friedelite, hedyphane, sphalerite, calcite, feldspar	0.1–0.2
Zincite, galena, cyprene, franklinite, schefferite, fowlerite, calcium larsenite, calcium rhodonite, calcite, biotite, sussexite	0.0–0.1

tivity from weakly active silicates such as sussexite (0.007) to apatite (7) and thorite (about 50,000) distinguish these deposits from most others studied in this series in which the heaviest elements are distributed sparsely and nearly uniformly. This distribution of radioactivity is doubtless related to the mode of origin, since concentration of radioactivity in the later pegmatitic stages of differentiation of granitic magmas is well known. In the present instance pegmatitic minerals were generally found to be more radioactive.

The distribution of radioactivity between the minerals may be partly related to period of deposition, younger minerals being more active, but generally seems to have been controlled by chemical affinities, those minerals which are most active being those which are frequently associated with radioactive elements. The lowest activities for any particular mineral was found in the purest crystals; complex aggregates tended to be higher.

Coeur d'Alene District:

Considerable variation was found in the activities of galena ores from mines in the Coeur d'Alene district, Idaho (0.2–2.7). Specimens from the Morningstar, Sunshine, and Tamarack veins were the least radioactive, and impure galena ores were relatively high. A small decrease in the radioactivity of country rock away from the ore suggests that the ore-bearing solutions were more radioactive than country rock. Lead and zinc ore with quartz—1.4; wall rock, 2 feet from ore—1.25; wall rock, 10 feet from vein 1.00. The results summarized in Table 5 show that sulphide ores are usually lower than country rock, but note below that some relatively radioactive sulphides were found at Gilman, Colorado.

Gilman, Colorado:

A wide range of radioactivities was found for rocks and minerals in the Gilman area (0.2 to 98.6), the distribution of radioactivity in ore minerals being approximately: sphalerite and galena, Eagle Mine—0.45 (3 specimens); sphalerite and pyrite—4.8 (1); pyrite 1.4 (2); chalcopyrite 2.2 (1); beegerite, hessite, chalcopyrite and pyrite, pocket in the 20th level (possibly radioactively contaminated) 98.6. The results show that some of the ore-bearing solutions in this region must have been appreciably radioactive; a group of 15 mineralized rock samples averaged 5, and covered a range from 2 to 8 alphas/mg./hr. Another group of 19 mineralized rocks was much less radioactive (0.7) as was much of the lead and zinc ore. The average activities for the sulphides as a whole, like those in the Coeur d'Alene district, were higher than the average of all sulphides from various localities. This suggests that the radioactivity of ore solutions of the Eastern Cordilleran metallogenetic province was relatively high, (as also indicated by the presence of radioactive ores in the district) and that conclusion is strengthened by the relatively high radioactivity of granitic and monozonitic intrusives in this region (Keevil in preparation).

Radioactivities of rocks in the Gilman area are summarized in Table 4.

TABLE 4. RADIOACTIVITIES OF ROCKS AT THE EAGLE MINE, COLORADO

alphas/mg./hr.

		Locality			
Rock	*Geological Age*	*Newhouse 16-2 and #2 Surface and Underground Sect.*		*19-5 and 20-1, 28, 36 Sections Included Radon**	
Weber limestone	Permo-Pennsylvanian	3.0 (1)			
Weber arkose	Permo-Pennsylvanian	1.6 (2)			
Quartz monzonite porphyry	Pennsylvanian	3.0 (3)			
Zebra dolomitic limestone	Upper Leadville Mississippian	0–8 (4)		0.26 (7)	0.44 (7)
Lower Leadville limestone	Mississippian	5.6 (1)	1.0 (8)		
Parting dolomitic quartzite	Devonian	0.4 (4)		1.00 (17)	0.95 (11)
Blue dolomitic limestone	Devonian	6.3 (1)	0.8 (7)	0.41 (13)	0.65 (17)
Granular sandstone	Ordovician	0.9 (3)			
Suwatch dolomitic sandstone	Cambrian	1.0 (3)			
Purple quartzite	Cambrian	0.06–4.3 (2)			
Mineralized quartzite	Cambrian	0.15–4.8 (7)			
Unmineralized quartzite	Cambrian	0.4 (2)			
Porphyritic biotite diorite	pre-Cambrian	1.15 (2)			
Mineralized and sanded rock		4.7 (3)		0.71 (18)	1.15 (19)
Fault breccia		0.0 (1)		0.47 (9)	0.64 (8)
Miscellaneous rocks mostly blue limestone		—		0.58 (36)	

* Radon concentrations in arbitrary units of alpha-rays from radon observed per gram of sample.

The rocks in the Newhouse sections showed greater variability than usual, the effect of ore-bearing solutions apparently being to increase the radioactivity of host rock above the uniform levels of radioactivity at the time of deposition. This suggested that a radioactive metamorphic aureole might be present around the ore deposits in this locality. A detailed study of the 19-5 level, however, did not show any systematic change of radioactivity along the sections, and only a few of the mineralized samples showed small increases in radioactivity above the average radioactivity.

As a part of this investigation one hundred determinations of the radon loosely held by the sample were made about two weeks after collection of the specimens.

It will be noticed in Table 4 that the average radon concentration (shown in final column) parallels the activities, and as expected, a decrease of radon with decreasing porosity is also suggested by the results, less radon being contained by less porous quartzite. Several points where relatively high concentrations of radon were found coincided with relatively high activities of the rock, suggesting local diffusion of radon from radioactive grains; however, other high radon concentrations were associated with porous "sand" zones and fault breccias which may have been due to concentration of radon by diffusion along associated fault structures and storage in beds which acted as better reservoirs for gases. The relatively low radioactivities of ore in this particular section would seem to preclude association of these anomalies with radioactive mineralization, particularly as the activities were often low for the solid sample when occluded radon gas was high.

Two typical series of results (for specimens taken at 25 foot intervals) along the sections are as follows:

	Activity	*Radon*		*Activity*	*Radon*
Zebra limestone	0.25	0.20	Fault breccia	0.45	1.31
"Sand" dikes	0.6	1.68	Sand dike	0.4	1.34
"Sand" dikes	0.3	1.14	Blue limestone	0.4	0.42
Zebra limestone	0.25	0.82	Blue limestone	0.25	0.48
Mineralized fault breccia	0.95	1.18	Fault gouge	1.00	1.38
"Sand" mineralized pocket	0.55	0.36	Mineralized fault gouge	0.35	2.31
Parting quartzite	1.2	0.69	Pyritized "sanded" rock	—	1.03
Parting quartzite	2.0	1.72	Blue limestone	0.1	0.43
Sand pocket	0.3	0.33	Blue limestone	0.4	0.5
Parting quartzite	1.2	0.60	Blue limestone	0.1	0.2

Superior, Arizona:

In the Magma Mine, Arizona, the ores cover an average range of activity for sulphide minerals from 0.03 to 1.5 alphas/mg./hr. This is about the same as the span of activities for country rock. A suggestion of a rhythmic distribution of radioactivity was noticed in the diabase which may be related to ore-bearing solutions, but is more likely due to physical-chemical equilibria in the consolidation of the diabase.

Austinville, Virginia:

Several series of samples of Cambro-Silurian limestones and dolomites representing sections through and near the Malden and Fisher Field ore-bodies at Austinville, Virginia, were studied, radioactivity determinations being made on 62 specimens. Minor differences in activity were found between several dolomitic limestone members: Lower Adl gray dolomite 0.17; fine grained Ribbon limestone 0.32; mottled gray Rome dolomitic limestone 0.31; Malden gray limestone and dolomite 0.25; carbide limestone 0.21; Adl gray dolomite 0.31; crude ore 0.16; mineralized contact 0.52.

As in the Couer d'Alene district, the impure mineralized border of the ore zone is somewhat higher, suggesting that ore bearing solutions were more radioactive than host rock, although the ore itself is weakly radioactive. The results in sequences for seven sections away from ore bodies were as follows:

.27, .25, .26, .27, (.61) .35, *.43*, .25, .18, .19, .14, .05, .27, .03, (.38)
.37, .50, .47, *.76*, .68, .62, .20, .20, .20, .20, (.52)
ore, .10, .49, .30, .11, .04, .02 (.36)
.21, .26, .20, .16, .20, .16, .63, (over ore) .18, .20, (.08) .19
.37, .37, .50, *.41*, .24
(.20), .01, 15 (over ore) .12, .16, .02
.15, (.07) *.19*, (.00) .11

The italicized values represent rock mineralized with sulphide, and the bracketed values indicate anomalous values in the series.The latter sometimes coincide with a change in the type of limestone, and in this connection it was noticed that the lighter gray specimens of dolomite and limestone are generally less radioactive than the darker varieties. In a few specimens the presence of chert and silica are associated with slightly greater activity, while some weathered specimens are quite low in radioactivity. The presence of mineralization is definitely associated with slightly increased activity, but since the mineralization is also related to fault structures it should be borne in mind that radioactivity might have been introduced by diffusion or penetration of relatively radioactive solutions along faults.

Other districts:

The results of radioactivity determinations of ore and country rock in several mining districts are summarized in Table 5.

In general gold ores are only feebly radioactive. Base metal sulphide and oxide ores show some regional variation but are generally equally or less radioactive than country rock. Pure minerals are likely to be less active than impure ore, and greater activity is occasionally associated with mineralized rock, suggesting that ore-bearing solutions sometimes contain more radioactive elements than host rock. This is to be expected when there is a close relationship of the ore to pegmatites, but not with

TABLE 5. RADIOACTIVITY OF ORE AND COUNTRY ROCK

District	*Activity of Country Rock*	*Type of Ore*	*Activity of Ore*
Porcupine area, E. Ont.	0.42 (22)	gold quartz	0.15 (2)
Red Lake, Ont.	0.9 (15)	gold quartz	0.0 (1)
Sullivan Mine, Ont.	0.5 (94)	quartz vein	0.00
Cournor Mine, Que.	0.5 (94)	gold-pyrite	0.00
Sudbury, Ont.	2.7 (3)	sulphides	0.9 (4)
Michipicoten Harbour, Ont.	1.4 (1)	molybdenite	0.3 (1)
Sault Ste. Marie, Ont.	—	molybdenite	1.1 (1)
Buchans, Newfoundland	0.3–2.3 (21)	sulphides	0.3–2.2 (6)
Brittania, B. C.	0.3–1.1 (6)	sulphides and schist	0.0–0.5 (3)
Jefferson City, Tenn.	0.3–3.2 (4)	sulphides with dolomite	0.3–2.7 (6)
Hanover, N. M.	—	zinc-lead ore	0.15 (4)
Mexico	1.15 (7)	sphalerite and pyrite	0.2 (1)
Coeur d'Alene district, Idaho	1.37 (4)	galena and quartz	0.4 (6)
Coeur d'Alene district, Idaho	—	galena and country rock	1.6 (5)
Magma Mine, Ariz.	0.51 (33)	sulphides	0.5 (6)
Austinville, Va.	0.26 (60)	sulphides	0.15 (2)
Gilman, Colo.	0.0–8.0 (107)	sulphides	0.2–2.2 (7)
Gilman, Colo.	—	iron, silver, copper, sulphides	98 (?) (1)
North of Great Slave Lake, N.W.T.	0.8–7.2 (3)	magnetite & sulphides	14 (1)
Baire Oriente, Cuba	9.5 (1)	pyrolusite ore	1.8 (1)
Franklin, N. J.	0.2–44 (55)	complex group	0.7 (91)
North of L. Huron, Ont.	—	iron oxides	1.46 (7)
Sterrett Mine, E. Que.	0.1 (2)	chromite	0.00 (1)
Kilmar, Que.	0.3 (7)	magnesite	0.01 (2)
Ivigtut, Greenland	0.1–5 (3)	cryolite and sphalerite	0.01 (3)

later veins, which appear to be generally impoverished in thorium and uranium.

DISTRIBUTION OF RADIOACTIVITY AROUND RADIOACTIVE ORE BODIES

Colorado:

In the metallogenetic province of eastern Colorado radioactive minerals frequently occur in pegmatites, and uranium ore is mined on a small scale in several localities. Although no detailed work was done on specimens from the immediate vicinity of such mines, the granitic rocks in the region are distinctly more radioactive than normal, and radioactive ore might have concentrated from magmas which gave rise to these parent rocks.

Aside from relatively higher activities of the Silver Plume batholith near mineralized zones, the regional differences in radioactivities found for two hundred Colorado granite specimens are due largely to differences in time of emplacement; the economic significance of the data is merely in the suggestion that parent rocks to be favorable for radioactive ore occurrence only have to reach in the magma stage, a certain minimum concentration level of radioactivity.

Great Bear Lake:

More than one hundred specimens of rocks from the vicinity of the Eldorado Mine on the southeast shore of Great Bear Lake, N.W.T., have been studied. Specimens of ore, of course, have an activity of several thousand, and a few specimens of highly altered rocks near veins were also found to be highly radioactive. In general, however, the rocks in this region are not much more radioactive than normal, the activities of specimens collected underground, a few dozen feet from ore, not being significantly different from the average for similar type rocks from more distant parts of the area.

Representative specimens of all the important types of rock near Eldorado Mine were collected through the kindness of Mr. Gilbert LaBine and Mr. E. J. Bolger. The sampling locations were distributed as uniformly as possible on the surface within several hundred feet of the coastline near LaBine Point and for some distance north and northeast beyond Glacier Lake, and also underground at fairly regular intervals, chiefly along the 500 foot and 650 foot levels, except for a group close to the number 3 vein where several rock types were sampled. Since no significant areal distribution of radioactivity was noted around the veins, a map showing exact locations is not given.

The results obtained for all except a few highly radioactive mineralized and highly altered specimens are summarized in Table 6. The radioactivities are generally higher than normal, but a surprising degree of uniformity was found in the results considering the danger of contamination by minute particles of radioactive minerals during handling. One would also have thought that the solutions would have penetrated greater distances away from the ore, relative to the zone of alteration often observed around mineral deposits. At the Eldorado Mine, high activities (exceeding 10,000) occur in the vein material and fall several thousandfold approaching normal radioactivities within a few feet of the vein, suggesting small permeability of country rocks to the radioactive ore components. As a matter of fact, no differences were found between rocks near the mine and those miles away except in a few cases where extensive alteration or mineralization was observed. The results show that geophysical prospecting for radioactive veins of this type, by means of field counting instruments, could be used for direct location of radiation, or

TABLE 6. RADIOACTIVITIES OF ROCKS NEAR LABINE POINT AND IN THE ELDORADO MINE, N.W.T.

Type of Rock	*No. of Specs.*	*Radioactivity Range Observed*	*Average Activity*	*Average for Other N.W.T. Specs.*	*Continental Average for Corresponding Rocks*
		alphas/mg./hr.			
Old complex (sed. gneiss)	3	2.30–2.87	2.49		
Greenstone	2	1.50–1.63	1.56		
Sediments	23	1.24–4.47*	2.60		1.60
Vein breccia	1	—	1.87		
Chloritic alteration & sheared rock	5	2.33–6.05	3.97		
"Red alteration"	2	2.03–4.46	3.25		
Porphyry	23	1.13–3.24	2.15		0.90
Granodiorite	8	0.86–4.07	2.43	1.09 (6)	1.75
Granite	7	2.34–5.23	3.48	2.45 (81)	2.02
Aplite	2	5.40–6.23	5.82		3.20
Diabase	6	0.57–1.90	1.07	0.12 (7)	0.72

* One value 7.07.

emanation from radioactive veins, but would not be guided by a larger surrounding aureole of higher than average radioactivities of country rock except in a vague or erratic way. In glaciated regions, a guide to ore might be provided by higher soil activities in the direction of glacial movement away from radioactive ore bodies. This subject will be discussed more fully in a later section.

Aplitic rocks, probably representing a late stage of the intrusion of granite, are the most radioactive rocks at Eldorado, and often elsewhere. The sequence of radioactivities down to the least active basic rocks is normal, and this suggests that the normal course of differentiation was followed as far as the radioactive elements are concerned. Activities for granite and aplite of the same order of magnitude have been found in other localities, so that localization of radioactive ore at Great Bear Lake is likely due as much to favorable sedimentary host rocks and favorable structural channels for the passage and deposition of radioactive ore, as to the geochemical nature of parent magma.

Possibilities of Radioactive Ore in Other Areas:

Radioactivities of granitic rocks, higher than the parent body at Great Bear Lake, have been found in wide areas in West Central Alaska, North Central British Columbia, Colorado, Texas, Georgia and New Hampshire, (and locally in a number of other scattered localities in the

United States and Canada). As the magmas producing these rocks might produce radioactive ore solutions under appropriate conditions, it might be well to examine parts of these areas with radioactive geophysical methods, and to search for favorable structure conditions and suitable host rocks.

Detection of Hidden Radioactive Deposits and Structures

Radioactive Ore:

Deposits containing appreciable quantities of uranium or thorium minerals are thousands of times more radioactive than ordinary rocks and mineral deposits, pure uranium or thorium oxides having an activity of 300,000, a monazite sand about 20,000 and radioactive zircon grains up to 15,000 alphas/mg./hr., as compared to a value of 2 for typical granitic rocks.

In considering the problem of locating hidden radioactive ore bodies from their characteristic radiations, it is important to realize that even the most penetrating gamma radiation is absorbed by a few feet of rock or overburden. The depths at which deposits are detectable can be increased through the use of more sensitive Geiger counters, or scintillation counters. Results with a Geiger counter with a sensitivity more than a hundred times that of the usual field counter are shown in Fig. 3. The usefulness of such instruments is increased by indirect radiation from material transported from the source. This includes the diffusion of radioactive emanation from the source, transport of radioactive material by solution, tracing the mother lode from active alluvial, detrital, glacial or secondary residual deposits, and geological guesses about the relationships of ore deposits to radioactive host rocks and structures.

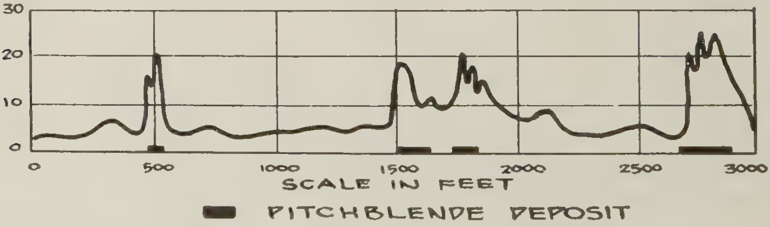

FIG. 3. Radioactivity profile from super Geiger counter survey over the Camray property, Alona Bay District, Sault Ste. Marie, Ontario. (Units are relative to a standard radium source equivalent to 3×10^{-9} R).

Diffused radon:

The only appreciably long-lived gaseous member among all the series of radioactive elements is radon, and this has a half-life of 3.84 days, decaying spontaneously to 1 per cent of its original concentration in 3 weeks. In the past numerous attempts have been made to employ radon in geophysical exploration for the location of fault structures and mineral deposits with varying degrees of success. The writer has obtained slightly higher concentrations of radon over a fault in the Gulf Coast, as well as a slight increase in radon concentrations with depth.* Higher radon concentrations were also found over a shallow oil field in Metcalfe, Township, Ontario.

In spite of the short half-life of radon, there is enough favorable evidence to warrant exploring the method further in the location of shallow fault structures, particularly where it is likely that relatively radioactive sources at depth are present to provide a reasonable gradient.

Migrating Solutions:

The upward transport of radioactive material from the source would tend to counteract the disadvantages of the short half-life of radon, provided the radioactive members (uranium, ionium and particularly radium) parent to radon, were included in the migrating fluids.

The radioactivity might be measurable in springs or ground water, or from residual matter along faults or in detrital material and this might lead to the discovery of radioactive ore below. Professor Esper S. Larsen has noted the occurrence of radioactive minerals in springs, and Lane and Bennett have located a fault in Michigan by the locally high radioactivity of water. Occasionally contacts between formations show slight increases in radioactivity as contrasted with that of either formation suggesting that fluids, migrating along the path of least resistance, have deposited small quantities of radioactive material.

Glacial Transportation of Radioactive Material:

Glacial movement over radioactive ore would provide glacial drift in the direction of movement, with specific radioactivities higher than the regional average for normal drift. In order to test this possibility soil and rock samples were collected in the vicinity of Wilberforce, Ontario, and south to Lake Ontario. These samples were collected in connection with a field trip made for other purposes, and it is unfortunate that the deposits around Wilberforce turned out to be much less radioactive than commonly thought. There are some small pockets of pegmatite containing uranite, and some syenite bodies of appreciable size which are considerably more radioactive than regional granites and gneisses. The results showed a tendency for higher soil activities in a cone representing the direction of glacial transport.

Further tests in more radioactive areas should be made to test the possibility of radioactive soil surveys as a reconnaissance method of exploration for radioactive ore.

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* Soil radon concentrations corrected for contributions of radon from mineral constituents. Average radon de-emanation of 8 to 10% was found in over 100 soil samples.

WELL LOGGING

Correlating strata by means of their characteristic natural radioactivities is widely used in logging oil wells, particularly where the penetrating power of gamma-rays permits logging through cased wells. Radioactivity produced in the more common elements by bombardment with neutron sources also has been employed to log wells with similar counting devices. Determinations of the absolute radioactivity of samples taken from wells and comparisons with the gamma-ray logs is needed to improve calibration and interpretation.

The writer has made a few alpha-ray logs of oil and gas wells; the results of which are available upon request.

ORE-GENESIS

In the family of hydrothermal ores, lead sulphides are made up partially of radiogenic material. An average of 9.8 per cent of the lead is derived from the decay of uranium and thorium during the period before segregation of lead ore. Isotopic analyses of lead ores have shown that the average lead-uranium ratio in the source material was variable, covering the range 3.2 to 15.0. This compares with the lead-uranium ratios in granites and intermediate rocks exposed on the earth's surface (probable value 9 ∓ 4). With these facts in mind and remembering that average thorium-uranium ratios of rocks lie within the range 3 to 4, it is a simple matter to show that the atomic weight of rock lead would be close to 207.21. The range in atomic weights actually found for ore-leads fall within a narrow range entirely consistent with what would be expected for their derivation from rocks or rock magmas. As a matter of fact, the change in isotopic composition of ore-lead with geological time and the variations in composition observed, suggest that hydrothermal ores of this type were derived from sources like rock magmas whose characteristics change with time, and not from a deep-seated uniform sub-stratum as proposed by Holmes on the basis of earlier radioactivity and isotopic data.

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THE MARYSVILLE GRANODIORITE STOCK, MONTANA

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ABSTRACT

The Marysville granodiorite stock of Montana is six miles north of the northern end of the Boulder bathylith. Around the stock cluster the rich gold-bearing veins that made the Marysville mining district famous in the early history of Montana. The Marysville stock is only one of many stocks that occur in the zone surrounding the large Boulder bathylith. The matter of determining whether these bodies are genetically linked to the bathylith is an elusive problem often met in areal geology. Three hypotheses have been advanced to account for the relationship between the Boulder bathylith, the Marysville stock, and the gold veins geographically associated with the stock: (1) the stock is an outlier of the bathylith and the veins are genetically related to the stock; (2) the stock, though an outlier of the bathylith, did not bring the ore but by its position determined the location of the veins by structural control; (3) the stock is much younger than the bathylith, is Oligocene in fact, instead of late Cretaceous or late Paleocene and was the gold-bringer. The second interpretation appears to be greatly strengthened by new evidence obtained during recent detailed investigation of the region by the writer, and the conclusion is reached that the Marysville stock is an outlier, probably a cupola of the Boulder bathylith, as is indicated or proved by the numerous scheelite-bearing pyrometasomatic deposits formed by the stock in its contact aureole as well as by the bathylith in its aureole, and that the gold veins were formed much later, in connection with the dacite extrusions just south of Marysville which are auriferous, or with the still later rhyolite eruptions, which also are auriferous.

The Marysville stock is surrounded by late Precambrian rocks of the Belt series—in ascending order, the Empire formation, the Helena limestone, renamed here Helena dolomite, the Marsh formation, and the Greenhorn Mountain quartzite, a newly recognized formation. The stock has profoundly metamorphosed the Empire formation and the Helena dolomite through which it cuts. The metamorphism of the Helena "limestone" was described by Barrell in his classic memoir on the district, but not that of the Empire formation, which was transformed into striking cordierite hornfelses. Scheelitic, scapolitic, vesuvianitic, and cordieritic metamorphism not previously recorded are here briefly described.

INTRODUCTION

During recent years the writer, assisted by Eleanora B. Knopf, has been mapping the northern portion of the Boulder bathylith of Montana and the rocks that make up its framework. The Boulder bathylith, though generally considered to be well known, has never been mapped petrographically. It is shown on the various maps so far published as quartz monzonite "qm," which procedure has been a notable labor-saving device. The bathylith in reality is composite, consisting of granodiorite, quartz monzonite, tonalite, white granite, muscovite granite, alaskite, aplite, graphophyre, diorite, and gabbro. During the course of this recent work the Marysville stock, its contact aureole, and the surrounding rocks were examined. Many stocks occur around the border of the Boulder bathylith, but the Marysville stock is best known, because of the important precious metal mining district geographically associated with it and because of Barrell's classic memoir on the district, in which he independently developed the hypothesis of magmatic stoping and made a penetrating study of the contact metamorphism produced by the stock. The relationship between Boulder bathylith, Marysville stock, and the geographically associated gold-silver veins has given rise to three hypotheses, one of which appears to be greatly strengthened by newly found evidence. The rocks surrounding the Marysville stock belong to the Belt series, in fact the type section of the Empire "shale" is within the district; these Beltian formations have necessarily been examined in detail. The Empire has never been described before; the Helena "limestone" is shown to consist mainly of dolomite and is therefore here renamed the Helena dolomite; the Marsh "shale," said for 50 years to be 300 feet thick, proves to be a formation 3000 feet thick; and a newly recognized Beltian formation, the Greenhorn Mountain quartzite, which conformably overlies the Marsh formation is described.

General Geologic Setting

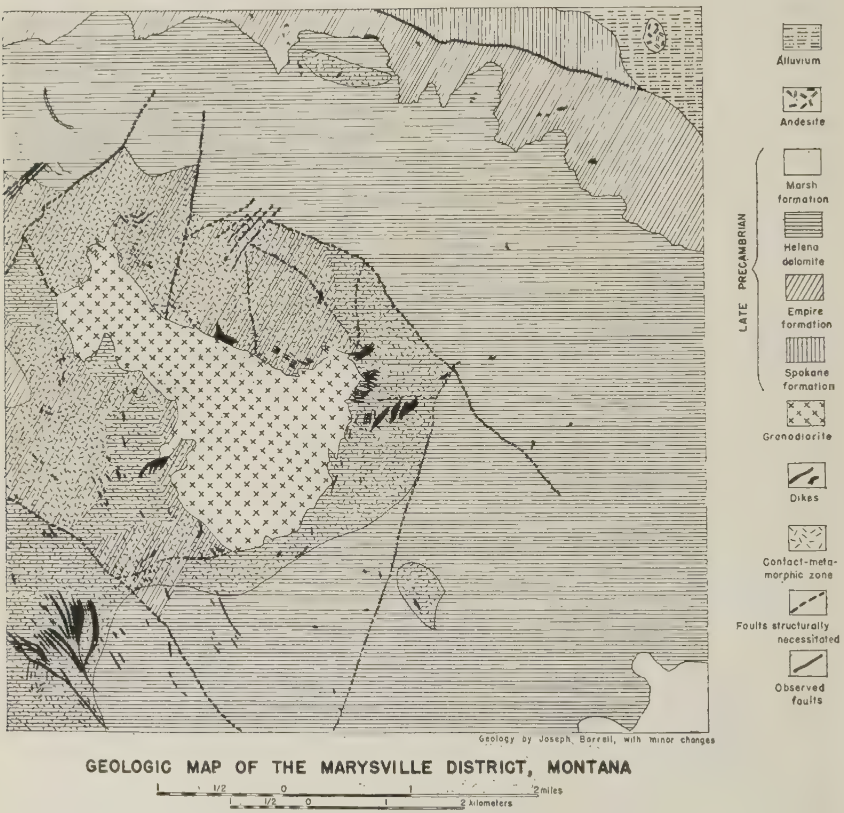
The northern end of the Boulder bathylith is on the Continental Divide just south of Mullan Pass. Six miles north is the Marysville stock, a body of granodiorite of 3 square miles surface exposure, which is intrusive into the Precambrian Beltian formations—the Empire formation and the Helena dolomite. In view of its small size the stock has produced a remarkable contact aureole one-half to 2 miles wide, consisting mainly of banded calcic hornfelses (Fig. 1).

The town of Marysville, now a ghost of its former self, is 20 miles northwest of Helena, with which it is connected by a first-class highway.

Empire Formation

The Empire formation, or "shale" as Walcott (1899, p. 207) called it, was named from its exposure west of the Empire mine, at the head of Lost Horse Gulch in the western part of the Marysville district. The lowermost beds exposed here are intruded by the granodiorite of the Marysville stock. All rocks in the type section are hornfelsed, dark-weathering, and are crackle breccias ramified with tremolite veinlets and quartz veins. They are well within the contact-metamorphic aureole of the Marysville stock and are overlain conformably by white diopsidic hornfels ($d=3.01$) of the Helena dolomite. Walcott mentioned that the Empire shale is well shown also on the summit of Drumlummon Hill, in the eastern part of the Marysville district, but the rock there proves to be a superb scapolite-diopside hornfels ($d=3.10$).

Barrell (1907) in his classic account of the Marysville district, ignored the type section of the Empire and described the formation as exposed on Long Creek in the north-central part of the Marysville area, outside of the contact aureole, where "it consists of finely laminated, soft limy shale, grayish green or buff-colored, with a few reddish bands." As seen by the writer, the Empire there consists of pale greenish hard argillite ($d=2.67$) alternating with mud-cracked rose or lavender argillites. The top of the Empire is prevailingly light-colored and weathers pale buff and yellowish. The buff-weathering habit of the uppermost part of the formation necessitates special care in distinguishing the Empire from the overlying Helena dolomite. As the section in this locality is bottomed by a fault, Barrell could measure only 520 feet in the Empire and accepted Walcott's figure of 600 feet as the total thickness of the Empire. However, the base of Walcott's section, as already mentioned, is an intrusive contact.



Geology by Joseph Barrell, with minor changes

FIG. 1

According to Pardee (1933, p. 126) the Empire consists mainly of siliceous greenish-gray shale. As measured in White and Avalanche gulches in the Belt Mountains, where the stratigraphic relations are well shown, it "appears to be about 1000 feet thick."

The Empire formation was found during the present work to be best exposed in a section that extends east-west through U.S.G.S. bench mark 4450 in the Scratchgravel Hills, just northwest of Helena. It consists of pale-green argillite, deep-red (brick-red, maroon, and lavender) argillites identical in appearance with those in the underlying Spokane formation, and fine-grained light-green and white quartzite. The thickness of the Empire is about 1000 feet.

The quartzite beds, which have not been mentioned in previous descriptions of the Empire, distinguish the formation from the underlying Spokane. The Empire overlies the Spokane conformably and the first appearance of quartzite in the section is taken to mark the bottom of the Empire. The lower quartzites are feldspathic rocks ($d=2.64$) and weather tawny, whereas those from the top, like the associated argillite, are buff-weathering. The transition to the overlying Helena dolomite is abrupt, taking place within a stratigraphic thickness of 20 feet.

The Empire formation is hornfelsed in much of the region here considered. Dark cordierite hornfels is common, as befits the general pelitic nature of the Empire formation, but interbeds of white calc-hornfels occur. Notable are the coarse cordieritic hornfelses on the north side of Ottawa Gulch in the Marysville district; large fine outcrops show well-developed cross-lamination extending across beds 6 to 12 inches thick, the cordierites being localized along the diagonal structures.

HELENA DOLOMITE

The Helena dolomite as it is here renamed was originally called the Helena limestone by Walcott (1899, p. 207); it consists largely of buff-weathering dolomite. This buff-weathering feature has been mentioned by all later investigators as typical of the formation, but no one has mentioned its dolomitic composition. The rock is dark gray on fresh fracture, is aphanitic, is non-effervescent on being touched with cold, dilute HCl, is compact and non-porous, and ranges in density from 2.76 to 2.82. Silica, as kindly determined by Dr. George Switzer on typical material ($d=2.78$), runs 32.6 per cent.

The Helena dolomite is transformed around the border of the Boulder bathylith and the Marysville stock into wide aureoles of white and cream-colored diopsidic and tremolitic hornfelses; in other words, into calcic and highly magnesian hornfelses. In fundamental studies of contact metamorphism it is of course important to know the composition of the unaltered rock, but in the Marysville memoir the Helena dolomite is described as an impure limestone. Harker and Marr (1893) appear to be the only geologists who have recorded that after finding abundant calcium-magnesium silicates in the specimens collected from metamorphosed limestone they went back into the field and re-examined the unmetamorphosed beds and found them to be dolomitic.

The formation contains subordinate limestone, which weathers gray or "bluish" gray, some siliceous limestone oolite, and beds of edgewise conglomerate up to 2 feet thick. Particularly notable are the *Collenia* biostromes, up to 10 feet thick, that occur at intervals from bottom to top of the formation; they have not been previously recorded as occurring in the Helena dolomite. In the Marysville district a large *Collenia* biostrome is exposed near Sawmill Creek on the abandoned right-of-way of the Northern Pacific Railway.

The thickness of the Helena "limestone" at Marysville was tentatively estimated by Barrell to be 4000 feet, and recent measurement along what appears to be a more favorable section south of the district confirms that figure.

MARSH FORMATION

The Marsh formation overlies the Helena dolomite conformably, as is well shown in excellent exposures on South Park Street in the heart of Helena. Here it consists of maroon mud-cracked shale (or argillite) and thin interbeds of mud-flake breccia and quartzite. It is only 250 feet thick here and is overlain disconformably by the Middle Cambrian Flathead quartzite.

The type section of the Marsh formation is at the mouth of Marsh Creek just outside of the Marysville map-area. The stratigraphic relations are not well shown there however, and Barrell ignored the type section. According to Walcott (1899, p. 208) the Marsh formation at the type locality is 300 feet thick and constitutes the topmost portion of the Belt terrane and this characterization, though wholly inadequate, has now done duty for 50 years.

By mapping the Marsh formation from Helena northwestward to Marysville, it is found to increase in thickness to 3000 feet and to be overlain conformably by a great thickness of quartzite, here named the Greenhorn Mountain quartzite. The top of the Marsh formation on the north flank of Greenhorn Mountain is a deep-maroon argillite, ripple-marked and mud-cracked and interbedded with thin quartzites 1 to 2 inches thick.

Deep red argillites, maroonish, purplish and violet, are common constituents of the Marsh, but quartzites in belts as much as 500 feet thick, cross-laminated siltstones, and laminites carrying casts of salt crystals are interbedded with them. The red argillites can not be distinguished from the similar argillites prominent in the Empire and Spokane formations. Probably the most distinctive lithic member in the Marsh formation is the dark red quartzite that weathers to a deep blackish red.

Greenhorn Mountain Quartzite

Above the Marsh formation and below the Flathead quartzite on Greenhorn Mountain is 1800 feet of quartzite. This figure is a minimum, as the top of the section is bounded by the erosion surface on which rests the Flathead quartzite of Middle Cambrian age. The pre-Flathead quartzite constitutes the country rock of the Continental Divide, making up the massive bulk of Greenhorn Mountain, 7500 feet in altitude, which forms the highest summit between Mullan Pass and Marysville. This newly recognized formation of the Belt group is therefore called the Greenhorn Mountain quartzite. The quartzite is notably feldspathic, the feldspar consisting of clear limpid microcline. In this feature the Greenhorn Mountain quartzite differs essentially from the Flathead quartzite, which is pure quartzite. The lower portion of the formation is massive quartzite, in places showing festoon cross-lamination, whereas the upper third is well stratified in beds of uniform thickness ranging from 1 to 2 inches.

In extant geologic maps the Greenhorn Mountain quartzite has been included with "undifferentiated Devonian and Cambrian" strata, with the result that the Flathead, which averages 100 feet in thickness with little deviation from this thickness a over distance of scores of miles, is implied to have a thickness on Greenhorn Mountain of several thousand feet.

Dacite and Rhyolite

Subsequent to the uncovering of the Boulder bathylith, dacite was erupted in large volume in the region between Marysville and Butte. Significantly the dacite encloses auriferous veins, proving that a gold metallization took place long after the great metallization produced by the Boulder bathylith.

A striking rhyolite, white and dotted with phenocrysts of coal-black quartz, occupies much of the area west and southwest of Marysville. It overlies the Oligocene clays near Mullan Pass, so is younger than the clays and judged from its relation to the present topography it is probably late Tertiary. Somewhat south of Mullan Pass, in the Rimini district, contemporaneous rhyolites are auriferous. Thus, in the general region of the Boulder bathylith there were at least three periods of gold metallization.

GRANODIORITE OF THE MARYSVILLE STOCK

The Marysville stock is three square miles in extent. It cuts through the Empire formation and Helena dolomite, which it has profoundly metamorphosed. Its emplacement according to Barrell was effected chiefly by rifting off of large angular blocks and their subsequent engulfment in the magma, a process, as is well known, termed by Daly magmatic stoping.

The main rock of the stock is a normal medium-grained granodiorite. It was called "quartz diorite" by Barrell, but the type material contains 15.6 per cent of orthoclase, and "granodiorite" is more appropriate, as pointed out by Lindgren (1907). Granodiorite ($d=2.75$) as obtained in ideally fresh condition from a newly blasted cut on the main highway just below the Drumlummon mill, is of typical appearance, the only exceptional feature being the presence of orthoclase in poikilitic patches 15 mm. in diameter. The rock consists dominantly of a zoned plagioclase ranging from An_{40} to An_{15} in the peripheral zone, quartz, orthoclase, hornblende, and biotite; it has an ideal hypidiomorphic granular fabric. Closely similar rock carrying even more conspicuous poikilitic orthoclase occurs on Drinkwater Creek two miles northwest; both were taken from the marginal zone of the stock. Two specimens ($d=2.68$), taken by the writer from the interior of the stock carry less hornblende and biotite (about 5 per cent) and may indicate, as suggested by Cross (Barrell, p. 55), that a younger intrusion is present but this problem was not investigated.

CONTACT METAMORPHISM

General features.—The granodiorite is surrounded by a remarkable contact-metamorphic aureole one-half to two miles wide. Within this aureole the Empire formation has been transformed into dark cordierite hornfels and the Helena dolomite into white and other light-colored diopsidic and tremolitic hornfelses. The contact zone was carefully mapped by Barrell; the calcic hornfelses were the main concern during his penetrating investigation, and the cordierite hornfels of the Empire formation was not mentioned. In that study he drew a sharp distinction between simple contact metamorphism (purely thermal, without addition of new material) and contact metasomatism, during which new material was added to form the contact rock. This distinction was made later, in 1911, by V. M. Goldschmidt in his masterly treatise on contact-metamorphism in the Oslo district, Norway, and the distinction is now taken for granted. Barrell found that where pure thermal metamorphism and metasomatic effects occur together, the metasomatic alteration is demonstrably the younger and has followed fractures in the already thermally metamorphosed rocks. Tremolite and diopside were found to be the common minerals in the contact aureole. Later work has shown that scapolite is widespread throughout the contact zone, that vesuvianite tactites have formed, that the Empire formation was changed to striking cordierite hornfelses, and that scheelite was widely introduced by late pyrometasomatic activity.

Cordierite Hornfelses of the Empire Formation.—The Empire formation is well exposed on the north side of the upper part of Ottawa Gulch; it is here horizontal and highly metamorphosed. Some of the rocks are cordierite hornfels in which the cordierites on weathered surfaces resemble small fusulines; in other beds they are large, from ½ to ¾ inch long. Large fine outcrops of hornfels show much cross-lamination extending across strata 6 to 12 inches thick; the cordierites are localized along the diagonal laminae. The hornfels ($d=2.69$) proves under the microscope to be made up essentially of quartz, microperthite, biotite, muscovite, and porphyroblastic cordierite notable for its brilliant yellow pleochroic halos.

Calcic and Magnesian Hornfelses of the Helena Dolomite.—Within the contact zone shown in Fig. 1 the Helena dolomite has been converted into compact (non-porous) hornfelses, white, light-gray, and creamy in color. Most are extremely fine-grained, generally aphanitic but in places ranging to microcrystalline. They range in density from 2.81 to 3.15, the lighter rocks proving to consist chiefly of tremolite and the heavier of diopside.

The hornfelses are particularly well exposed along the highway on the north side of Silver Creek, where they abut upon the stock. Beds consisting wholly of tremolite hornfels ($d=2.91$) are interstratified with those made up wholly of diopside ($d=3.01$). Composition rather than proximity to the granodiorite contact appears to have been the factor controlling whether tremolite or diopside have formed,

Rarely are the minerals of the hornfelses megascopically recognizable. An exception is the vesuvianitic diopside hornfels ($d=3.12$) from the north end of the ridge between Drinkwater and Deer creeks; in this layered hornfels vesuvianite in euhedral crystals is localized along certain beds ¾ of an inch thick.

Because of the minute grain of the hornfelses every thin section is likely to afford a surprise. The first section examined during the present study was cut from the remarkable banded hornfels in Trinity Gulch. The hornfels proved to be made up of diopside and abundant spongiform scapolite, a mineral not previously recorded from the Marysville district. Later, superb scapolite-diopside hornfels was found to occur on the summit of Drumlummon Hill. The rock ($d=3.10$) is white, coarse-grained, and consists largely of diopside prisms 15 mm. long, which are idiomorphic against scapolite. The scapolite amounts to 10 per cent; its indices —$N_z=1.56$ and $N_x=1.542$—indicate a scapolite containing 55 per cent of the marialite molecule. Minor constituents of the hornfels are calcite, tremolite, phlogopite, muscovite, and seybertite, which has developed at the expense of the diopside.

Scheelite-bearing Contact Rocks.—As the result of the announcement by the writer in the "Helena Independent" on Sept. 26, 1941, that scheelite had been found in a garnet contact deposit on the margin of the Boulder bathylith west of Helena, scheelite was soon found throughout a wide region in contact zones, in old mines and prospects, and in the clean-up of the gold dredge operating on Prickly Pear Creek. Scheelite was found in the placers of Piegan Gulch in the northern part of the Marysville district, and soon after was found in place in many localities in the contact zone of that district. Additional finds were made by the writer in the summer of 1942, and these confirmed that scheelite had been introduced into the Marysville contact aureole at many places.

The scheelite occurs in two ways:

(1) With garnet as thin stratiform layers, which represent limestone beds in the Helena dolomite that had escaped hornfelsing and were later replaced by andradite garnet and scheelite; and (2) belts of calcic hornfels that have been permeated and replaced by epidote, garnet, amphibole, quartz, and scheelite. Typically, epidote predominates, and consequently the scheelitic tactites are dark-green rocks. Although some good scheelite ore has been found, the deposits so far known have not proved to be of economic value.

The principal development of scheelite-bearing garnet-epidote rock occurs on the ridge between Deer and Drinkwater creeks. Here in a belt of white calcic hornfelses 3200 feet wide are 30 scheelite-bearing zones ranging from 10 inches to 3 feet in thickness. The rocks of this belt were mapped by Barrell as Empire, but the prevalence of white calcic hornfelses, some of which show "molar-tooth" structure, proves that the section consists of hornfelsed Helena dolomite. The garnetiferous beds were briefly discussed by Barrell (Barrell, p. 134), but in 1901 no one knew that scheelite may occur as a contact-metasomatic mineral. These scheelitic tactites are somewhat unusual in that they have been formed by the garnetization and epidotization of thin-bedded diopside hornfelses.

Three stratiform layers of scheelitic tactite were found in a stratigraphic interval of 200 feet on the north flank of Mount Belmont. The most promising is 2 feet thick. The ore consists of epidote, quartz, amphibole, calcite, and scheelite.

The Gold Veins.—The veins formerly so productive yielded gold and silver. Their general features, including their remarkable lamellar quartz and calcite gangue, were described earlier (Knopf, pp. 64–68) and indicate, contrary to the interpretations of Weed and Barrell, that they are not genetically allied to the Marysville stock.

The geographic association of the veins and the stock appears to have been determined by structural control. Bruce (1937, p. 378) has pointed out that the proximity of ore bodies to exposed igneous rocks has been commonly determined by "structural factors only and has no genetic significance."

If the gold-silver veins of the Marysville district are genetically related to the intrusion of the granodiorite stock, scheelite could be expected to occur in the veins, in view of the widespread introduction of scheelite into the contact-metamorphic aureole.

During the present investigation some thousands of feet of the workings of the Drumlummon mine, the principal mine in the district, were rayed with an ultra-violet lamp, but no scheelite was found. This result was not unexpected however, as the scheelite deposits are pyrometasomatic and the veins are clearly of epithermal type in the classification of Lindgren (1933, pp. 450, 463).

Relationship of Marysville Stock, Gold Veins, and Boulder Bathylith.—Three solutions of this problem have been given, in 1907, 1913, and 1933. According to Barrell (1907) the Marysville stock, lying 6 miles north of the northern end of the Boulder bathylith, is an outlier of the bathylith. The chief evidence cited in support of this idea is that the Marysville granodiorite is petrographically and chemically like that of the rock of the bathylith as represented at Butte (Barrell, pp. 24, 55). However, this evidence is not necessarily conclusive, especially in view of the composite nature of the Boulder bathylith, which ranges from tonalite through granodiorite, quartz monzonite, and granite to alaskite. The associated precious-metal veins were thought to have been produced by contraction effects on the margin of the stock and to have been filled by mineral matter deposited from emanations released from the solidifying magma.

In 1913 the idea was advanced that though the Marysville stock is an outlier of the Boulder bathylith, the geographically associated gold-silver veins are not genetically allied to the stock (Knopf, pp. 62–68). The veins were shown to differ greatly in their principal features from the deposits genetically allied to the bathylith. They are essentially unlike in metal content, in persistence of ore in depth, and in the remarkable lamellar or platy nature of their quartz-calcite gangue. They are in fact identical in their chief properties with the precious-metal veins that are inclosed in the dacite lava piles, which rest on the eroded surface of the bathylith.

In 1933 a new solution was offered by Pardee and Schrader (1933, pp. 20, 29, 69). They accepted the idea that the Marysville veins were formed during an epoch of ore deposition that was later than the metallogenesis accompanying the cooling of the Boulder bathylith, but they maintained that the veins are genetically related to the Marysville stock.

Consequently the stock is considered to be younger than the Boulder bathylith [the date of whose emplacement has not yet been more closely established than "late Cretaceous-early Tertiary"] and they considered that the age of the stock is probably Oligocene.

The recent discovery that scheelite has been widely introduced into the contact zone of the Marysville stock greatly strengthens the conclusion that the Marysville stock, as Barrell first pointed out, is an outlier of the bathylith. Whether the "outlier" is a cupola on the bathylith or is a crosscutting stock is not determinable. Further evidence that the two igneous masses are essentially of the same age is that the contact-metamorphic effects produced by bathylith and stock on the Helena dolomite are identical both in intensity and in extent.

The epithermal nature of the gold-silver veins indicates, as it did in 1913, that the precious metal deposition is younger than the metallogenesis of the Boulder bathylith. This conclusion has gained further strength from the fact, which was not known in 1913, that gold occurs in many places in the extensive dacite-covered area a few miles south of Marysville. As the dacites, which are supposedly late Miocene in age, rest in places on the eroded surface of the bathylith, a long span of time must have elapsed between the intrusion of the Marysville stock and the subsequent deposition of the gold-silver veins associated with it.

Nearly fifty years have been required to formulate a reasonable answer to the problem of the relationship between Boulder bathylith, Marysville stock, and the geographically associated gold-silver veins; perhaps when, as physicists promise us, five grams of a rock will suffice for a determination of its absolute age, the answer will come more rapidly and more positively.

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LEAD-URANIUM RATIO AND POSSIBLE GEOLOGIC AGE OF ALLANITE FROM GREENWICH, MASSACHUSETTS[1]

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ABSTRACT

Allanite from a pegmatitic lens in the Monson granodiorite at Greenwich, Massachusetts, contains 0.03% Pb, 1.53% Th, 0.095% U. The results of other studies are outlined, and a maximum possible age of about 390 million years is calculated.

On May 27, 1942 the writer received from Dr. Robert Balk, at that time Chairman of the Department of Geology, Mount Holyoke College, a number of small chips of allanite, containing some small quartz inclusions, marked "Allanite crystals, from pegmatitic lens, in injection gneiss (Emerson's Monson gneiss, *U.S.G.S.*, *Bull. 597*), $\frac{1}{2}$ mile E. of Greenwich Village, Belchertown quadr., Mass. Collected by R. Balk." The material had to be put aside until after the war. In 1946, on learning that the locality was no longer accessible due to the flooding of the Quabbin Reservoir, and that in any event there was no more fresh allanite there, it was decided to undertake examination of the sample, though the small size made it extremely doubtful if enough lead could be extracted for determination of the isotope ratios. So far as is now possible, the work has been completed, and it seems worth while to put the results on record, even though a complete age determination appears impossible.

The allanite was found by Balk in the course of prolonged field studies in the area subsequently flooded. His report is at present in manuscript in the hands of the U. S. Geological Survey. He describes the locality in the following terms (Balk, 1949):

> "Abandoned quarry, west base of hill 751 ft., SE corner of the new $7\frac{1}{2}$-minute Quabbin Reservoir Quadrangle, Massachusetts. This hill is now an island in the reservoir. Before the area was flooded, and the new $7\frac{1}{2}$-minute quadrangles were published, the locality was: W base of hill 760+ft., $\frac{1}{4}$-mile due east of Greenwich village, 15-minute Belchertown Quadrangle, Massachusetts."

The gneiss in which occurs the pegmatitic lens containing the allanite, forms part of the Monson granodiorite of Emerson (1917), who mapped it as late Carboniferous. His age assignment is probably made on the basis of the Carboniferous age given to the Worcester phyllite some 30-odd miles to the east, but this is not specifically stated in the paper. In his earlier work in this area, Emerson (1898) refers to this rock as the Monson gneiss, and lists it as Lower Cambrian, placing it above the supposed

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[1] Contribution from Committee on Measurement of Geologic Time. Division of Geology and Geography, National Research Council.

Algonkian rocks of the western part of the state. In his geological map of this date the area in question is apparently mapped as Cambrian Becket gneiss. On the other hand, Billings (1934, *et seq.*), and Chapman (1948) consider the Oliverian gneisses of New Hampshire as post-Silurian, and it may well be (though but little field work in the intervening areas has been done) that the Monson granodiorite is a southward continuation of the New Hampshire rocks. Knopf (1949) says that the Monson granodiorite has been traced southward into Connecticut and it may be that the Spinelli and Strickland quarries of the Middletown-Portland, Connecticut area are allied to it. Radioactive age determinations on minerals from these quarries (Holmes, 1947, pp. 136–7) yield a figure of approximately 260 million years. This, on the basis of Holmes' latest views (1947), represents the lower Mississippian. The Monson granodiorite is definitely older than the Triassic rocks of the Connecticut Valley to the west. It has been suggested, though the writer can find no references in published works, that the rocks of the Central Massachusetts Upland are very much older than those of the Worcester Trough to the east. Whether or not this "very much older" means pre-Cambrian or early Paleozoic was never made clear, but presumably the latter was meant. There has always been a tendency to call, on first examination, any highly metamorphosed or extremely gneissoid rock of uncertain origin pre-Cambrian until evidence to the contrary is forthcoming. As regards the rocks of this area, this view apparently was not held long enough to get into print.

It thus appears that the field evidence as to the relative age of the Monson granodiorite is somewhat contradictory. The field relations are complex throughout the New England area, and are especially so in the Central Massachusetts Upland. Thus any evidence as to the absolute age, even though incomplete, will be of some interest. As will appear later, the present work has by no means resolved the differences of opinion.

The optical properties of this sample of allanite were studied, in a very preliminary fashion, by the writer, who found virtually all of the fragments examined to be anisotropic. As this looked hopeful—isotropic allanite is metamict, and nearly always gives meaningless lead ratios because of alteration—several randomly selected chips were given to Miss Jewell J. Glass of the U.S. Geological Survey, who reported as follows (1947):

> "Allanite, from Greenwich, Massachusetts. Small fragments of broken crystals or crystalline material. Color jet black with pitchy luster in hand specimen, brownish smoky-gray in thin grains. No cleavage observed. Fuses with swelling. Gelatinizes with hot HCl.
>
> "Optical properties: In grain mounts in immersion oils the crushed powder of a single grain varies from completely isotropic to weakly anisotropic. Isotropic grains smoky gray. Index variable, $n=1.698$–1.702. Anisotropic grains feebly pleochroic brownish gray to

greenish gray. Absorption: $Z>Y>X$. Negative, $2V$ large. Indices variable. Average indices: $\alpha=1.706$, $\beta=1.711$, $\gamma=1.714$. Associated with quartz and biotite."

An autoradiograph of this material, enlarged 5×, made by Dr. Herman Yagoda of the National Institutes of Health, appears as Fig. 1. The chips of allanite, etc., were mounted in bakelite, by the technique usual for making polished sections, and then the surface ground flat. The original exposure was 933 hours. In the print the quartz inclusions appear as small black areas randomly scattered through the allanite. Dr. Yagoda points out (1946) that the seven chips apparently are of very nearly equal radioactivity, and says further that "Microscopic examination of the polished sample reveals the presence of minute white inclusions"—the quartz inclusions noted above—and that "radiocolloid structures are absent from these samples." Radiocolloids are considered to indicate incipient alteration or the interruption of radioactive equilibrium by some outside process.

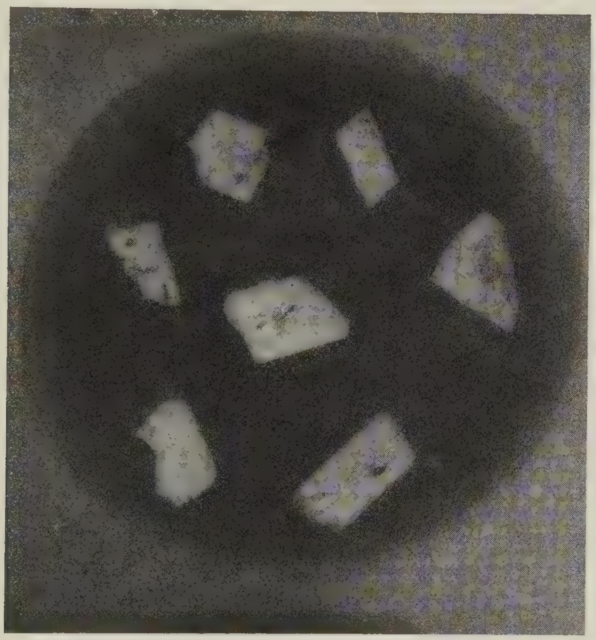

FIG. 1

Dr. Yagoda has calculated the total surface activity, i.e. the activity due to U+Th for each unit of surface, from the writer's analysis,[2] and also from examination of the autoradiographs (Yagoda, 1949, p. 164). The agreement is as good as for any other material so studied.

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[2] The analytical figures quoted by Yagoda in the reference given, were supplied by the writer on the basis of early calculations based on one analysis. The difference from those given in the present paper is insignificant for the purposes for which Yagoda used them.

All of the available sample, except for the chips used by Miss Glass and by Dr. Yagoda, and three or four deposited as a reference sample in the U.S. National Museum, was used for the analyses. The material was crushed in a clean steel mortar, mulled to a medium powder with a clean steel roller and plate, and the quartz with a few small fragments of biotite separated with fresh, washed bromoform. Separation was aided by the greater brittleness of the allanite, which caused it to be in somewhat finer powder than the light minerals, and so they tended to float. Two treatments produced light and heavy fractions that showed no mutual contamination when examined under the binocular microscope. The allanite fraction was thoroughly washed with acetone, dried on the steam-bath, and then allowed to remain exposed to dust-free air for several days to ensure complete removal of organic matter.

The analyses were carried out by methods outlined by the writer in several previous papers and in several Reports of the Committee on Measurement of Geologic Time. For the "age analyses" redistilled acids and 3× recrystallized Na_2CO_3 were used for the solution of the sample and the determination of lead. The mineral decomposed readily in 1:1 HCl, but as usual, repeated evaporation and retreatment was needed to dehydrate the silica. Lead was first separated as the sulfide, and determined as the sulfate, with examination of all filtrates for possible traces of this element. Thorium was separated from uranium, etc., as the oxalate, separated from rare earths by repeated precipitation as peroxy-nitrate, and determined as the oxide. Iron, etc., were separated from uranium by precipitation with ammonia and ammonium carbonate, repeated as often as tests showed necessary. A very small "heavy metal" fraction (probably largely $V_2O_5+MoO_3$) was then scavenged with cupferron, uranium precipitated with carbonate-free redistilled ammonia, and weighed as U_3O_8.

Because of the very small total uranium it was not possible to separate it into UO_2 and UO_3 fractions. The large excess of iron, nearly $\frac{2}{3}$ of which is in the ferrous state, would presumably have resulted in the complete oxidation of the uranium. The separation of the rare earths into La and Y groups is approximate only, as usual, based on the relative solubility of their double potassium sulfates. It is felt that no further comments on analytical methods are needed.

As the lead content was small it was necessary to take as large samples as possible to make the accuracy of the determination of this element as great as could be. In Analysis II the combined solutions after the determination of lead were made up to a known volume, and four determina-

TABLE I. "AGE ANALYSIS" OF ALLANITE FROM GREENWICH, MASSACHUSETTS

% Pb		% Th		% U	
A_GI	0.031	A_GI	1.520	A_GII (a+b)	0.091
A_GII	0.033	A_GIIa	1.577	A_GII (c+d)	0.098
		A_GIIb	1.544		
		A_GIIc	1.578		
		A_GIId	1.430		
Ave.	=0.032	Ave.	=1.530	Ave.	=0.095

tions of thorium made on aliquot parts thereof. This was done partly for convenience in working with smaller quantities of major constituents, and partly to obtain further checks on the thorium content. After the removal of iron, etc., the first two aliquots were combined and one determination of uranium made on them, and the same procedure was followed for the last two. Due to an accident, the uranium fraction of Analysis I was lost. The above procedure accounts for the varying number of Pb, Th, and U determinations shown in Table I. The complete analysis of the mineral given in Table II is based on from 1 to 5 determinations of each constituent, made during the course of the various "age analyses". One determination of "total rare earths" gave 17.64%, which agrees reasonably well with the sum of Ce, La group, Y group oxides of 18.48%. From the reasonably close agreement of the summation to 100%, it appears that no constituent of importance was missed. In one case, the filtrate from a Pb determination was tested for Cu and other elements precipitated with H_2S in dilute acid solution, but no weighable amount was found. There appear to be no peculiarities of

TABLE II. COMPLETE ANALYSIS OF ALLANITE FROM GREENWICH, MASSACHUSETTS

SiO_2	32.22%	CaO	10.86
Pb	0.03	SrO	0.02
ThO_2	1.74	MgO	1.66
U_3O_8	0.12	Ce_2O_3	3.86
Fe_2O_3	4.87	La_2O_3, etc.	13.92
FeO	9.07	Y_2O_3, etc.	0.70
Al_2O_3	17.89	$MoO_3+V_2O_5$, etc.	0.03
TiO_2	0.29	MnO	0.15
K_2O	0.14	H_2O, +110°	1.81
Na_2O	0.29	H_2O, −110°	0.09
			Total=99.76%

composition to distinguish this material from any other "typical" allanite.

The water content is a little higher than is desirable in such material when used for age determination. This, plus the presence of some isotropic grains reported by Miss Glass, suggests incipient alteration. On the other hand, the high FeO content is reassuring. Tentatively we may classify this as "good grade B" material.

It should be borne in mind that, in summing the analysis of a mineral containing radiogenic lead the lead is substituting for uranium or thorium in an uranium oxide or thorium oxide lattice. Thus the lead is present as the element Pb, and not as the oxide PbO. When the quantity of lead is very small, as in this case, the effect on the summation is negligible, but when it is considerable, as in geologically old pitchblendes, the difference needs to be taken into account. In the calculations in these analyses the atomic weight of the lead was taken as 207.25, based on the Th/U ratio. Any other value in the vicinity of 207.2 would probably have served equally well.

From these analytical data we now come to calculate the approximate age. For a mineral of a high Th/U ratio such as this, the value of the constant "k" in the approximate age equation (Pb/U+kTh) becomes important. This value is still not known with desirable exactness. The usual value given is 0.36, but some workers prefer 0.34. Dr. A. F. Kovarik (letter from A. Knopf to J. P. Marble, October 21, 1949) calculates 0.333 for a mineral of mid-Paleozoic age. As the two isotopes of uranium have different decay rates, the value of this "constant" varies, in theory, with the age of the mineral under consideration. However, as the amount of the isotope U^{235} is so much smaller than that of U^{238}, it probably suffices to use only three different values, for very old, moderately old, and relatively young minerals. Work now in progress on the exact values of the half-lives of U^{235}, U^{238} and Th will presumably clear the matter up. To avoid using these various values, we may take ages from the curves drawn by Wickman (1943), who has constructed graphs converting analytical data into ages, based on the then current values for the radioactive constants. Lopez de Azcona (1948) has constructed similar curves. The various calculations are summarized in Table III.

TABLE III. APPROXIMATE AGE CALCULATIONS, GREENWICH, MASS., ALLANITE

1. Pb/U+0.36 Th×7600 m.y.=0.049×7600 m.y.=c.370 m.y.
2. Pb/U+0.34 Th×7600 m.y.=0.051×7600 m.y.=c.390 m.y.
3. Pb/U+0.333 Th×7600 m.y.=0.053=7600 m.y.=c.400 m.y.
4. Pb/U+Th=0.02; U/U+Th=0.06; age from curve=c.390 m.y.

Average=c.390 million years

As was said earlier, it is extremely unfortunate that it is not at present possible to determine the isotope ratios of the lead extracted from this allanite. The weight of $PbSO_4$ obtained in Analysis I was 0.00040 g., in Analysis II, 0.00263 g., a total of about 3 milligrams. In the process of further purification and conversion to iodide, further losses would occur, and less than 5 milligrams is, at the moment, impossible to use, unless one is willing to take the chance that the first run will come out all right. This material is being saved, and when accurate mass-spectroscopic work can be done on less than 2.5 mg. of lead salt, we hope to be able to work this up into suitable form.

Meanwhile it must be borne in mind that the approximate age derived on the basis of the above calculations simply represents a maximum value. The presence of any "common" lead will reduce the age, and since the percentage of lead is the smallest number in the ratio, the effect of any change in this will be at a maximum.

Based on the tables given by Holmes (1947), which are based on the best available evidence, it appears that 390 million years comes about between the Ordovician and the Silurian. The presence of a comparatively small amount of "common" lead would reduce this figure to one corresponding to a late Silurian age, which would suggest that the views connecting the Monson granodiorite with the Oliverian magma series of New Hampshire were correct. However, a somewhat greater amount of "common" lead, but not an impossibly large quantity thereof, would reduce the age still further, and we might presumably end up with a Lower Carboniferous age. The effect of analytical errors is, of course, unpredictable. With the small amount of material here available it is not possible to make enough analyses to see if there is any trend in the results. The writer can only say that his normal procedures seemed to work without difficulty, and that he tried to take all the needful precautions.

The results make a pre-Cambrian age untenable, and also seem to throw out the Cambrian age given in Emerson's earlier paper. The field evidence is quite clear that these rocks are pre-Triassic. The writer is not willing to go farther than to say that the present, admittedly incomplete, work indicates a middle to lower Paleozoic age for this allanite from the Monson granodiorite.

The writer takes this opportunity to thank Dr. Robert Balk for supplying the material, describing the locality, and assisting in the discussion of the age relations; Miss Jewell J. Glass for making the optical determinations; Dr. Herman Yagoda for making and interpreting the autoradiographs; Dr. Adolph Knopf for assistance in the discussion; and the Secretary of the Smithsonian Institution for making it possible

for the writer to carry out the analyses and other processes in the laboratories of the U. S. National Museum.

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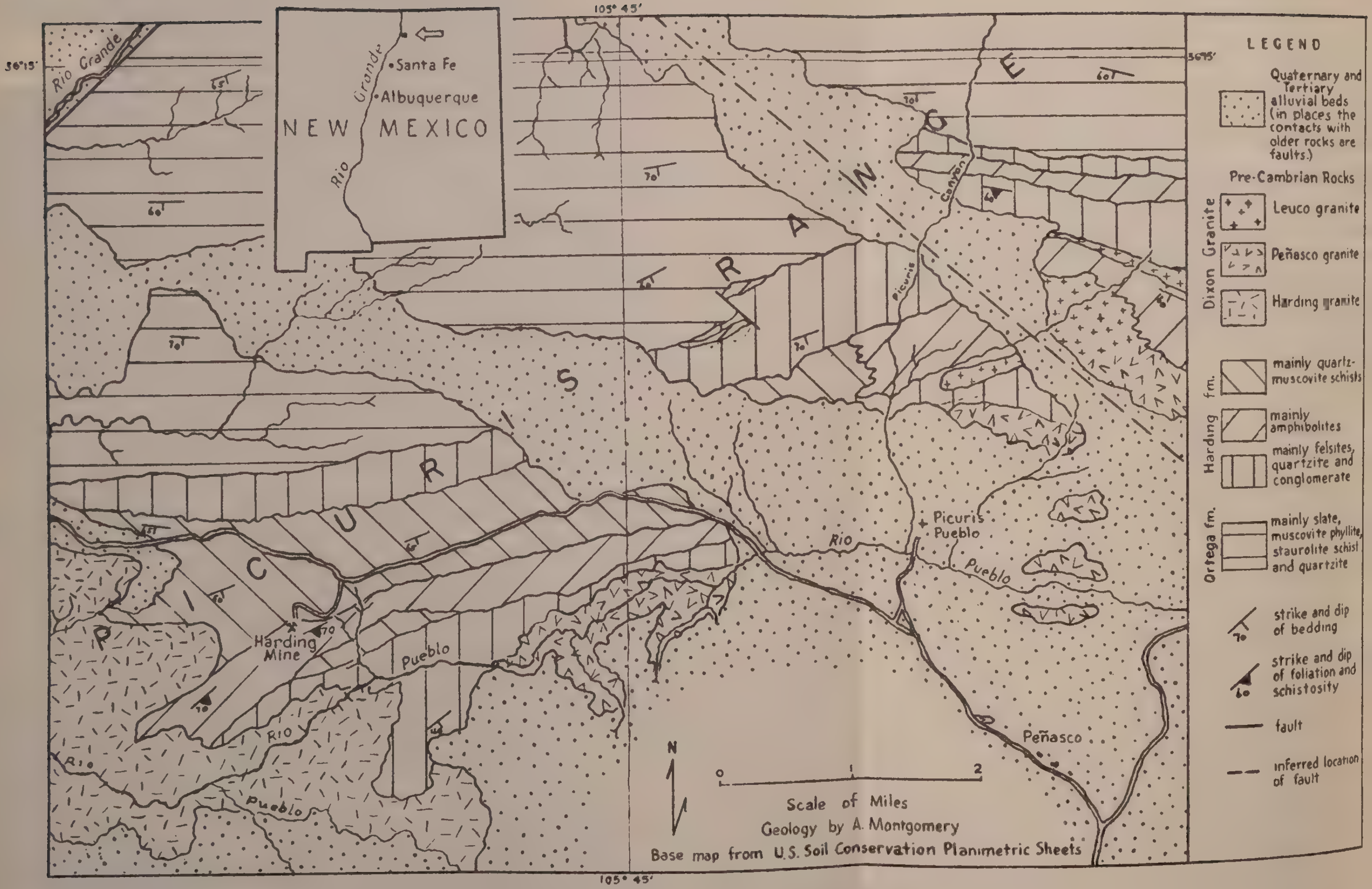

105° 45'
36°15'
36°15'
Rio Grande
Santa Fe
Albuquerque
Grande
Rio
NEW MEXICO
P I C U R I S
R A N G E
Picuris Canyon
Picuris Pueblo
Rio Pueblo
Harding Mine
Pueblo
Rio
Rio Pueblo
Peñasco
65
60
70
70
70
70
60
70
70
60
60
65
60
70
70
45
N
0 1 2
Scale of Miles
Geology by A. Montgomery
Base map from U.S. Soil Conservation Planimetric Sheets
LEGEND
Quaternary and Tertiary alluvial beds (in places the contacts with older rocks are faults.)
Pre-Cambrian Rocks
Dixon Granite
Leuco granite
Peñasco granite
Harding granite
Harding fm.
mainly quartz-muscovite schists
mainly amphibolites
mainly felsites, quartzite and conglomerate
Ortega fm.
mainly slate, muscovite phyllite, staurolite schist and quartzite
strike and dip of bedding
70
strike and dip of foliation and schistosity
60
fault
inferred location of fault
105° 45'

FIG. 1. Geologic map of the southwestern part of the Picuris Range, New Mexico.

GEOCHEMISTRY OF TANTALUM IN THE HARDING PEGMATITE, TAOS COUNTY, NEW MEXICO*

ARTHUR MONTGOMERY, *Harvard University, Cambridge, Massachusetts.*

ABSTRACT

The notable tantalum enrichment in the Harding pegmatite, in northern New Mexico, is believed due to (1) a parental tantalum-rich granite, (2) a uniquely favorable environment, (3) specific absence of tantalum-bearing sphene from all adjacent granite, (4) long-continued deformation acting upon crystallizing granite and pegmatite.

A more indirect factor, affecting the type of tantalum mineralogy peculiar to this pegmatite and perhaps responsible for a vertical mineralogical zoning, is believed to have been a chemical control exerted by pegmatite assimilation of country rock.

A complex series of geochemical events, all affecting each other, seems necessary for any such rare-element concentration.

INTRODUCTION

The Harding pegmatite is in north-central New Mexico, mid-way between Santa Fe and Taos and 5 miles due east of Dixon. It lies 20 miles from Taos, in the far southwest corner of the Picuris Range, a 16-mile-long spur projecting from the main north-south Sangre de Cristo Range. It was worked extensively in the 1920's for lepidolite, became the only important domestic source of tantalum during World War II, and now is one of the country's foremost beryl producers. The mineralogy of this pegmatite has been studied by Schaller and Henderson (1926), Jahns (*U.S. Geol. Survey report* in preparation) who mapped the deposit in 1942, and in greater detail in connection with drilling exploration by the U. S. Bureau of Mines in 1943 (*U.S. Bur. Mines*, 1943, 1944, 1946, 1949), and Montgomery (1950) who has done the operating for tantalum and beryl. The author owes his introduction to the Harding Mine to Prof. E. S. Larsen, Jr., who originally suggested to him that tantalum might be found there in quantity.

The pegmatite has intruded amphibolites and quartz-muscovite schists which form the major part of a younger pre-Cambrian rock group occupying the southern half of the Picuris Range. An older metasedimentary rock group occupies the northern half. The Dixon† granite (Just, 1937, pp. 24–25) borders the southern group along the southwest and southern range boundaries, also occupies the far-eastern end of the range. Pegmatites and quartz veins genetically related to the granite intrude granite and all older pre-Cambrian rocks. The granite lies only a half-mile south and west of the Harding pegmatite; to the west it cuts across bedding of schists and amphibolites.

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* Contribution from the Department of Mineralogy and Petrography, Harvard University, No. 318.

† This name has been changed to *Embudo.* In addition to this change, the name *Harding formation* should be changed to *Vadito* formation in Fig. 2.

DESCRIPTION OF THE PEGMATITE

The main pegmatite mass is an almost flat-lying lens that dips gently to the southwest, with an axial dip length of 1,000 feet, a strike width of 500 to 600 feet, and an average thickness of about 60 feet.

From the largest central mass exposed in the quarry, drilling has explored the pegmatite down-dip along its axial length for a distance of 900 feet. All of this part of the pegmatite is hidden beneath overlying amphibolites which increase in thickness from 10 to 20 feet at the quarry to more than 100 feet at the holes drilled furthest to the southwest. Drilling 700 feet southwest of the quarry has shown the pegmatite there to be a narrowing pipe-like mass roughly 70 feet thick and 70 feet wide; further southwest it thins rapidly and may pinch out in that direction beyond 900 feet. Several holes 50 to 100 feet to the southeast show the pipe-like mass narrowed down to a 15 foot thick dike dipping 45 degrees southeast and nearly parallel to foliation in the amphibolites.

Out around the dumps by the quarry the northwest part of the pegmatite thins abruptly; only a few portions of it remain uneroded in that area. Southeast of the quarry the pegmatite is hidden beneath amphibolites.

The well-exposed eastern extension is very irregular, thinning and passing beneath the North Knob and narrowing southeastward from there to a thin, branching dike following steep bedding and foliation (N60E, 60–70SE); then, 800 feet from the quarry, it swells to a thick mass, before thinning again and continuing another 1,000 feet down into the Southeast Gulch. Gulch outcrops represent a 2 to 10 foot thick dike ascending from below. Exposures here show an irregular step-like pattern as the dike successively follows bedding and crosscuts it in a tortuous ascent from an elevation of 7,200 to 7,400 feet at the mine. Near the mine the horizontal crosscutting process became paramount resulting in the great flat-lying Harding mass. This is not bedding control, for various beds of amphibolite, schist and quartzite are crosscut. The northern, much-eroded part of the pegmatite chiefly crosscuts quartz-muscovite schist; the central mass and southern parts cut across dark amphibolites.

INTERNAL STRUCTURE AND MINERALOGY

The internal structure shows a horizontal layering, with a topmost 2 foot thick wall zone of coarsely-intergrown quartz and muscovite with fist-size spots of white beryl and patches of albitized microcline. Large crystals of blue apatite and columbite-tantalite (averaging 45% Ta_2O_5), also some coarse lepidolite, occur sparingly. Where beryl is now being mined at the quarry's west end, a solid, almost pure, 2 to 5 foot thickness

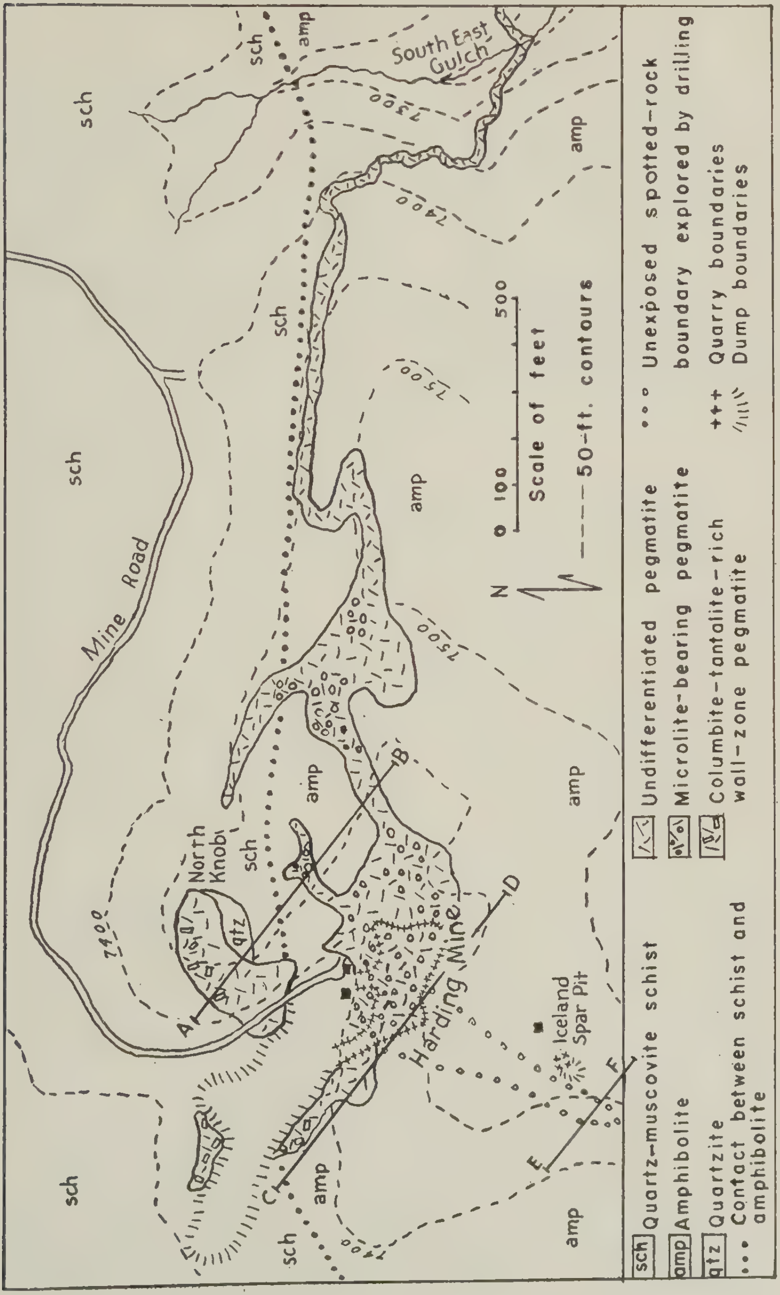
sch
amp
qtz
South East Gulch
7300
7400
7500
Mine Road
North Knob
Harding Mine
Iceland Spar Pit
A
B
C
D
E
F
N
0 100 500
Scale of feet
- - - 50-ft. contours
sch Quartz–muscovite schist
amp Amphibolite
qtz Quartzite
• • • Contact between schist and amphibolite
Undifferentiated pegmatite
Microlite-bearing pegmatite
Columbite-tantalite-rich wall-zone pegmatite
∘ ∘ ∘ Unexposed spotted-rock boundary explored by drilling
+++ Quarry boundaries
Dump boundaries

FIG. 2. Geologic map of the Harding pegmatite (based in part on mapping by U. S. Geological Survey).

of beryl occurs as the topmost layer; an albite-rich stringer in places comes down out of the amphibolite above to lie directly on top of the beryl as a thin border-zone layer. A 6 foot thickness of nearly pure quartz underlies the wall zone. Below this in the central quarry section is a 30 foot thick core zone of coarsely crystallized microcline, small rounded spodumene crystals, finely crystalline muscovite and lepidolite, and abundant minute crystals of columbite-tantalite and brownish-black microlite. At the top of this zone is a layer of rich lepidolite and coarse spodumene. This central spodumene-rich zone, called "spotted-rock" by Jahns, is the low-grade tantalum ore body developed down-dip for 650 feet by drilling. This zone ends abruptly southward against a southeasterly quartz zone 20 feet thick, full of huge, flat spodumene crystals crisscrossing each other like jackstraws and standing at steep angles to the horizontal. At the foot of this zone occur rich concentrations of spodumene and smoky

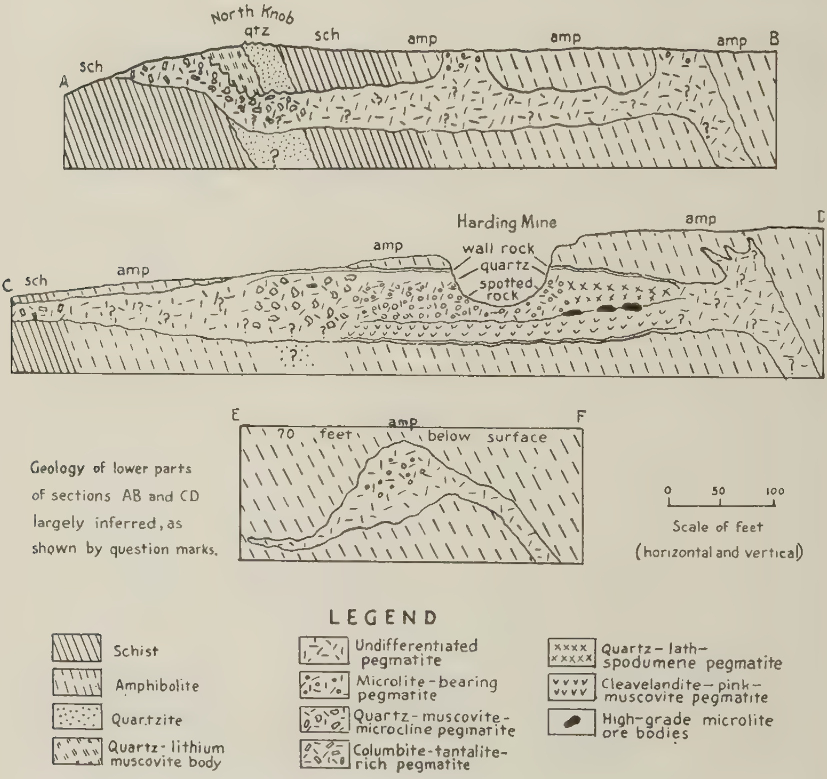
North Knob
qtz
sch
amp
amp
amp
B
A
sch
Harding Mine
amp
D
amp
wall rock
quartz
spotted rock
C
sch
amp
E
F
amp
70 feet below surface
Geology of lower parts of sections AB and CD largely inferred, as shown by question marks.
0
50
100
Scale of feet (horizontal and vertical)
LEGEND
Schist
Amphibolite
Quartzite
Quartz-lithium muscovite body
Undifferentiated pegmatite
Microlite-bearing pegmatite
Quartz-muscovite-microcline pegmatite
Columbite-tantalite-rich pegmatite
Quartz-lath-spodumene pegmatite
Cleavelandite-pink-muscovite pegmatite
High-grade microlite ore bodies

FIG. 3. Structure sections of the Harding pegmatite along *AB*, *CD*, and *EF*.

quartz, associated with large lepidolite lenses commonly averaging 10% in tiny yellow microlite crystals. This latest microlite ore mineralization is structurally controlled by fracturing and shearing. Below this ore mineralization is 20 to 30 feet of white cleavelandite with pink muscovite, largely hidden beneath the quarry floor but continuing northeastward under the "spotted rock" zone. The "spotted rock" ends abruptly northward at the northwest quarry portal against very coarse quartz-rich pegmatite containing blocky partly-albitized microcline and veined by patchy streaks of muscovite, gray lepidolite and smoky quartz. No spodumene or microlite occur in this distinctive northwest zone, but columbite-tantalite in coarse crystals becomes abundant in the overlying (largely-eroded) wall zone. Many hundreds of pounds of columbite-tantalite have been picked up in northern gulch placers. No microlite has been observed in these placers, either as loose grains or as inclusions in water-worn pieces of lepidolite.

The mineral paragenesis appears to show a long-continued hydrothermal history with later tantalum-lithium mineralization largely structurally controlled. Pre-mineral and post-mineral deformation are much in evidence.

A northwest-southeast cross-section through the pegmatite (Fig. 3, C-D) shows, then, not only a horizontal layering in part conforming to the expected internal structure of a flat-lying lens, but also a three-fold, vertical mineralogical zoning difficult to explain in terms of such structure.

Wall-Rock Control of Mineral Zoning

A possible explanation can be found for the vertical zoning if two factors which the author believes fundamental in the formation of this pegmatite are considered: (1) replacement of a large body of country rock, and (2) wall-rock control exerted by the different types of replaced country rock upon adjacent parts of the replacing, crystallizing pegmatite. Evidence for such a hypothesis comes from a study of the manner of occurrence of tantalum in this pegmatite.

On the North Knob above the quarry, prospect cuts have exposed a large body of fine-grained muscovite, lithium muscovite and quartz. This grades northward, and vertically downward also, into blocky quartz-microcline pegmatite similar to that at the northwest quarry portal. It grades southward into a mass of almost pure quartzite. All gradations may be seen in this area between quartzite or quartz schist and apparent pegmatite replacement of these. Possible replacement of unexposed quartzite might help to explain likewise the very coarse quartz-rich character of the northwest quarry-portal zone. A further significant fact is the abundance of calcium-free columbite-tantalite (and absence of calcium-rich microlite) in pegmatite adjacent to this zone; also the quartzite and quartz-muscovite schist replaced by most of this part of the pegmatite are free of calcium-bearing minerals.

The central and southeast pegmatite zones largely crosscut calcium-rich amphibolites; in these zones microlite occurs in great quantity and almost entirely takes the place of columbite-tantalite. The calcium from the assimilated amphibolites would go to make up the calcium tantalum oxide, microlite, $[(Na,Ca)_2Ta_2O_6(O,OH,F)]$. The quartz-rich character of the southeast zone could depend in part also on replacement of quartzite, since quartzite lenses outcrop in amphibolite along the southeastern pegmatite border. All of this is possible evidence favoring country rock control on tantalum mineralization of the pegmatite. Further possible evidence that calcium might have been set free in large quantity by pegmatite replacement of amphibolite derives from the rich iceland-spar occurrence (Kelley, 1940) 400 feet south of the quarry. A number of iceland-spar veins found elsewhere in the range all occur close to amphibolites much intruded by pegmatite.

In the Southern Rockies of New Mexico and Colorado are two other microlite-rich pegmatites, the Pidlite (Jahns, 1946) 40 miles south of the Harding Mine near Rociada and the Brown Derby (Hanley, 1946) 150 miles northward near Ohio City. Both occurrences have dark calcium-bearing wall rocks, amphibolite and meta-diorite. This is not to suggest that all microlite occurrences require calcium-rich wall rock, but the two remarkably similar regional occurrences mentioned at least do not disagree with the hypothesis advanced for the Harding pegmatite.

Geochemistry of Tantalum in the Pegmatite

Problems to be solved.—Why is there such a notable concentration of tantalum in the Harding pegmatite? Mining and drilling have proved a minimum total of 150 tons of the pure element. Together with the Brown Derby and Pidlite deposits, these three lithium-rich pegmatites are unique in this country for abundance of tantalum and manner of its occurrence in the rare tantalum mineral, microlite. One geochemical reason has already been suggested for the character and distribution of the tantalum minerals in the Harding; the still more fundamental problems relating to the source of this tantalum and the cause for its extreme concentration in this particular pegmatite will now be considered.

The Dixon granite a tantalum-rich granite.—Landes (1935) has described small regional pegmatite areas as characterized by distinctive rare-element or hydrothermal-mineral phases. A smiliar geochemical rare-element characterization applies over very great regional areas; ore-vein occurrences confirm this. Southern California pegmatites are thus boron-rich; in the same region are found many other boron minerals as axinite, dumortierite and borates. In the same way pegmatites over much of the eastern Southern Rockies are distinctly characterized by fluorine, the rare earths and their common associates, columbium and tantalum. From Denver southward pegmatite areas noted for such minerals as cerite, monazite, pyrochlore, bastnäsite, fergusonite, samarskite, columbite, etc., are legion. Monazite-rich pegmatites occur near Petaca, 30 miles northwest of the Harding Mine, and in the Pidlite pegmatite area 40 miles southward. Picuris pegmatites mostly lack such minerals, except for traces of tiny columbite crystals in some of them and the Harding columbite-tantalite and microlite.

From such evidence it seems valid to assume that in general "younger"[1] granitic magmas of the eastern Southern Rockies have been tantalum-rich. The Dixon granite, genetically related to the Harding pegmatite by field relations and by petrologic and mineralogical evidence,[2] represents one of these geochemically-related magmas.

The Harding pegmatite geochemical environment favorable for tantalum enrichment.—Rankama's (1944) classic research on Ta geochemistry shows that (1) this element is a typical associate of Cb and the rare earths, (2) in minerals it replaces most easily tetravalent Ti, Zr, Sn and W, and less easily, trivalent Cr and Mn, (3) at high temperatures it is most easily incorporated into the minerals, sphene, rutile, anatase, brookite, columbian rutile, cassiterite, wolframite, zircon, thorite, monazite and tourmaline, (4) it least easily enters the structures of apatite, garnet, beryl, iron oxides, sulfides, quartz and silicates in general (excluding titaniferous types, as biotite), (5) granites are far richer in it than other rocks, (6) it possesses a high energy of ionic migration, or high E-value (Wickman, 1944, p. 373), (7) it is not found in concentrations (as tantalum minerals) in early crystallized igneous rocks, but in those of late-magmatic stage, as pegmatites.

Points (6) and (7) appear to be anomalous, for with high energy necessary for ionic migration tantalum should be able to enter structures of high-temperature-formed minerals of early magmatic stage. Wickman (1944, p. 388) has answered this by suggesting that the minerals most apt to incorporate Ta at high temperature are so rare in early magmatic igneous rocks, that there is no favorable resting place for the element during that stage. It is thus left over for late-stage concentration. A normal granite with the usual high-temperature accessories, sphene, zircon and rutile, would house Ta in these minerals. If a crystallizing granite largely lacked such accessories, were also lean in biotite, and had been originally Ta-rich, then Ta conceivably might be left mostly free for abnormal concentration in later-formed genetically-related pegmatites. Such concentration would be further enhanced by a tendency for tantalum to form independent Ta-rich minerals alone, rather than to be dispersed as traces among other pegmatite minerals. This would be particularly true if such Ta-receptive hosts as monazite, titanium and rare-earth minerals, cassiterite, rutile, tourmaline, etc., were lacking in the pegmatite. Bray's (1942, p. 797) geochemical investigation of rocks of the Jamestown area, Colorado, found that pegmatite minerals there typically contained less of almost every rare element than the same minerals of related parent granites. His explanation was that the structures of the pegmatite minerals would be less open and less tolerant to rare-element substitution at their lower crystallization temperatures. In such a hypothetical case as outlined above, Ta might be left free in great quantity at a late stage of pegmatitic consolidation and be more or less forced to form crystallized compounds in its own right, provided the appropriate environment with a sufficiency of requisite chemical elements were available.

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[1] Late pre-Cambrian or younger.

[2] Discussed under *Dixon granite* and *Pegmatites* in Montgomery (1950).

The Ta geochemistry and mineral paragenesis of the Harding pegmatite perfectly fit the hypothetical case above. No Ta-receptive minerals are present in the pegmatite other than the independent Ta-rich minerals, columbite-tantalite and microlite. All the granite that borders the pegmatite on the south and west is unusually lean in sphene, zircon and rutile, and biotite is not abundant. Abundant Ca would have been available for late-stage microlite formation from pegmatitic replacement of amphibolite. Spectrochemical analysis shows the latest-formed, structurally controlled yellow microlite of the southeast zone to be purer and far richer in Ta than the earlier-formed black microlite and columbite-tantalite of the central and northeast mineral zones.

The immediate tantalum source during pegmatite formation.—The two sections above have brought out evidence for an original Ta-rich granitic magma in the Harding area and for that special geochemical environment favoring an exceptional Ta-enrichment in the Harding pegmatite. The final section deals with the most fundamental problem of all; namely, the reasons for the great and specific Ta source which must have been immediately available for this particular crystallizing pegmatite.

This study began three years ago when it was realized that significant geochemical clues bearing on the problem would depend on detailed geologic mapping backed by accurate spectrochemical analyses. Since then the geology of the area has been mapped (Montgomery, 1950) and the background lithology clarified; special attention was given to the Dixon granite. Dr. Harold C. Harrison, now at Rhode Island State College, cooperated in the study by setting up Ta standards and making many spectrochemical analyses. Much of the result depends upon his work.

Field work and spectrochemical research together were able to give both Rankama and Bray several highly significant geochemical correlations between certain rare-element-bearing pegmatites and quartz veins and genetically-related igneous rocks. Rankama (1944, p. 67) found in the Tammella, Finland, pre-Cambrian igneous complex that only the youngest granite contained Ta, and that it was this particular rock alone which could be related by field evidence to tantalum pegmatites of the area. Bray (1942, pp. 786–87) found in the Jamestown, Colorado, igneous complex that a cerium-rich granite could be related to cerite-bearing pegmatites of the area and that a vanadium-rich granodiorite probably was related to certain pyritic veins of the area containing the vanadium mica, roscoelite. It was hoped in the present study to find similar geochemical clues, especially since there was only one possible related igneous rock, the Dixon granite.

It was thought that the Dixon granite might contain sufficient Ta to be detectable by Harrison's spectrochemical analysis. Rankama (1944, p. 54) found an average Ta content of about .0004% for all granites investigated by him, and .0007% for the Ta-bearing younger ones. Several Ta-rich granites contained more than .001% Ta (Rankama, 1944, p. 36). The Dixon granite is a "younger" granite and has been dated in terms of the Harding pegmatite at about 800 million years (Ahrens, 1949, p. 252), or late pre-Cambrian.

Rankama's determinations of minute percentages of Ta depended upon *x*-ray spectrography and chemical pre-enrichment. Without such means, using the Harvard optical 2-meter grating spectrograph and a modified cathode-layer excitation technique, and finding the 2714 Å line to be the most sensitive, Harrison was unable to detect Ta in any Dixon granite samples. He was able to verify qualitative Ta estimates on sphene as low as .001–.002%. Sphene was selected because it had been found in quantity in certain parts of the Dixon granite. A number of sphenes from well-known localities had already been analyzed for Ta by Rankama (1944, p. 30) and von Hevesy (1929, p. 95). Sphenes in the Harvard mineral collection from identical localities were run by Harrison, and the tabulated results show good agreement (see Table 1). Sphene collected from Dixon granite several miles east of the Hardiny Mine showed a clearly detectable tantalum content of approximatelg .003%. Although the range of sensitivity used was too low for detection of Ta below .001%, this very fact minimized contamination problems and strengthened authenticity of results. Tantalum was sought in appreciable quantity, or not at all.

A study of the Dixon granite brought out two facts of great interest: all observed granite outcrops south and west of the mine contain traces of sphene and zircon, and no rutile; all observed granite outcrops from 2 miles southeast of the Harding Mine and for 6 miles eastward past the Picuris Pueblo show similar traces of zircon and no rutile, but also abundant sphene in the form of thinly disseminated flat crystals of average 1–3 mm. length. These last may be outlined by white leucoxene, and are easily seen with a hand lens. The sphene-bearing granite contains a minimum average of .3% sphene; in some areas it contains up to 4%. In this paper the darker biotite-rich sphene-bearing Dixon granite will be called the *Peñasco granite*, and the more metamorphosed, sphene-free Dixon granite south and west of the Harding Mine, the *Harding granite*. A third type of Dixon granite occurs further northeastward in the range, almost lacking in dark minerals, and is a leucogranite. The Peñasco granite is fresher and darker than the Harding granite. Study of the Dixon granite (Montgomery, 1950) shows this rock to be everywhere microcline-rich and of quartz monzonite character. It is high in silica and alkalies, and low in MgO and CaO. In most thin sections albite-oligoclase slightly exceeds microcline; orthoclase is absent. Porphyritic phases, with 10–20% of large microcline phenocrysts, occur near the Picuris Pueblo and west of the mine. Some of the more altered granite near the mine shows a small amount of muscovite of hydrothermal origin instead of biotite. Chemical analyses show the Peñasco granite to carry slightly more TiO_2 than the average Harding granite, as well as more Fe, Mg, Ca, but less silica. Differences appear largely due to differential assimilation effects. The importance of such assimilation is not known, but xenoliths of an inferred original dacitic

TABLE 1. Ta CONTENT OF TITANIUM MINERALS

	Rankama Ta_2O_5 %	von Hevesy Ta_2O_5 %	Harrison Ta %
Sphene, Eganville, Ontario	.001	.007	.001 (SiO_2 removed)
Sphene, Arendal, Norway		.003	.003–.005
Sphene (95% pure), Dixon granite, New Mexico			.003
Yttrotitanite, Kragerö, Norway		.01	.05–.08
Perovskite, Magnet Cove, Ark.	.01		(undetected)
Perovskite, Achmatowsk, Urals	.002	.0015	.001–.002

type are numerous all through the western area of the darker Peñasco granite; assimilation of such rock is believed responsible for the slightly more mafic character of the Peñasco granite. West of the Harding pegmatite most of the granite has cut across muscovite schists, and a few schistose inclusions have been found in it; assimilation of such rocks is believed partly responsible for its more felsic composition, lighter color and porphyritic character, although more severe metamorphism has also affected the resulting rock. Where amphibolites 1 mile southwest of the mine have been crosscut by granite, the granite ceases to be porphyritic and becomes much darker and biotite-rich; but it contains no sphene. Much meta-rhyolite and meta-quartz-latite are believed to have been assimilated by large masses of Harding granite which now appears as strongly metamorphosed granite, either light-colored or biotite-bearing, south of the mine and the Rio Pueblo.

Although differential assimilation effects thus furnish very important geochemical clues, they may not be responsible for the varying sphene content of the Harding and Peñasco granites. Darker biotite-rich granite southwest of the mine does not contain sphene. Comparison of chemical analyses of Peñasco and Harding granites (Montgomery, 1950) shows slightly less TiO_2 in the latter (.30% in the Harding granite (3 analyses), as against .46% for the Peñasco (2 analyses)). This difference is not believed sufficient to explain the absence of abundant sphene in the Harding granite; the difference is more likely related to the smaller biotite content of most of the Harding granite. Analysis of the darker biotite-rich Harding granite (not included in the TiO_2 percentages above) shows .55% TiO_2. It is possible that sphene crystallization in the Harding granite may have been prevented because of the more severe metamorphism suffered by it, perhaps during early stages of consolidation. This portion of the Dixon granite was more exposed to deformation and metamorphism because it apparently represents the earliest-crystallized, quickest-cooled outer shell of the intrusive. The sphene-bearing Peñasco granite is believed to represent a deeper-level, more protected portion of the intrusive.

Two points have been demonstrated: that it is all the same granite, and that near the mine the granite is sphene-free as compared to the sphene-rich Peñasco granite some distance removed. Assuming the Ta content to have been constant for the granitic magma giving rise to both Peñasco and Harding granite types, the absence of this sphene in the crystallizing Harding granite is believed to have left free the amount of Ta proved by spectrographic analysis as present in the sphene alone of the Peñasco granite. It would not require a very large body of crystallizing granite magma *without the sphene* to supply sufficient Ta from this one unlocked source for all the great tonnage of Ta minerals found within the Harding pegmatite. Based on the spectrochemical analyses, a square-mile of the Peñasco granite 1,000 feet thick contains 200 tons of Ta in its sphene alone.

ADDITIONAL FACTORS NEEDING CONSIDERATION

Rankama found .004% as an average Ta content for all Ta-bearing granites investigated by him. Only a tenth of this amount can be accounted for in the average Peñasco granite in terms of likely Ta contents for individual minerals present. The following rough estimates are largely based on Rankama's (1944, pp. 25–32) findings for the minerals concerned:

TABLE 2. ESTIMATED TOTAL Ta IN PEÑASCO GRANITE

	% of mineral in granite	Ta content of mineral in %	Ta content of granite in %
Coarse sphene (Harrison)	>.3	.003	.00001
Zircon, allanite, sphene, rutile (?), apatite	.01– .1	<.01	.00001
Magnetite, ilmenite (?), pyrite	.1 – .5	<.0001	.0000005
Epidote, sericite, kaolin	.5 – 1.0	<.00001	.0000001
Biotite (Harrison: <.001% Ta), muscovite (minor)	6.0 –10.0	<.0001	.00001
Albite-oligoclase	31.0	<.00001	.0000031
Microcline	26.0	<.00001	.0000026
Quartz	32.0	<.00001	.0000032
	Total Ta in granite		.0000395%

The discrepancy is large, even though it has been suggested (Rankama, priv. comm., 1950) that much of it may be due to large local variations of Ta content to be expected in granitic minerals. Rankama (1944, p. 52) anticipated some such possible anomalies by suggesting that some Ta might occur in Ta-rich rocks in the form of alkali tantalates as included substances. A small amount of undetectable Ta might be so dispersed but the relatively large quantity, which this study suggests may be missing, surely would be observable, recoverable and detectable regardless of its chemical state. The only way to clear up such a possible discrepancy would be to take such a Ta-rich granite as the Dixon granite is believed to be and, by means of *x*-ray spectrography and chemical pre-enrichment, to check and recheck the over-all Ta content, then to analyze for Ta each separated and purified mineral constituent in turn. In this way the whereabouts of any missing Ta should be discovered.

A final vital question must be answered: could the Harding Ta concentration be due to greatly increased volume of pegmatite alone? Rough estimates based on a careful mineralogical study of columbite-bearing Picuris pegmatites, and comparing one of the latter of typical outcrop size (200 feet long by 2 feet thick) and allowing for 1 inch of depth as surely exposed by weathered and broken pieces all along the outcrop, do not bear this out. Allowing for several ounces of columbite in the rock so exposed, a liberal allowance in terms of the very few minute columbites ever exposed along the whole of such an outcrop, a ratio of 1 lb. of Ta to 50--100 tons of pegmatite is the result. Mining and drilling have proved for the main present mass of the Harding pegmatite a minimum ratio of 1 pound of Ta to 8 tons of pegmatite. To explain the Harding Ta concentration, there would have to be nearly a ten-fold Ta increase above the maximum Ta content observed in columbite-bearing pegmatites of the area. A similar increase would be necessary for Be; for Li the increase would be fantastic. Why should the Harding body carry an extra abundance of these rare elements, far above all other Picuris pegmatites of similar and related type, because of greatly increased size alone? It is not necessary for a pegmatite rich in rare elements to have large size. Would not the simplest explanation be that the exceptional rare-element concentration found in the Harding depended upon an open system, fed during pegmatite consolidation by some sort of granitic or pegmatitic magma below? Much evidence explains the bulk of the Harding Ta-Be-Li enrichment as due to a long-continued process in which hydrothermal solutions from granitic magma below successively invaded a large partly-consolidated pegmatite mass undergoing severe structural deformation.

CONCLUSIONS

The exceptional Ta enrichment in the Harding pegmatite is believed due to these factors:

(1) a genetically related Ta-rich granite;

(2) some indirect wall-rock control which, during pegmatitic replacement of amphibolite, supplied to adjacent and replacing pegmatite the Ca needed for abundant formation of the rare Ta mineral, microlite;

(3) a uniquely favorable environment in which the parental crystallizing granite lacked the usual Ta-receptive minerals, and consolidating pegmatite itself lacked such minerals;

(4) specific absence of abundant Ta-bearing sphene, present in the same granite elsewhere, from all pegmatite-adjacent granite, which during crystallization of that granite may have freed, or left free, sufficient Ta from this source alone to supply the entire Harding enrichment;

(5) a long-continued process of deformation acting upon the consolidating pegmatite, fracturing it and keeping it open and accessible to successive invasions of hydrothermal solutions from granitic magma below which was under stress and being squeezed of its pegmatitic juices by great regional forces.

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SOME HORNFELSES FROM SAXONY AND THE PROBLEM OF METAMORPHIC FACIES

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The Mineralogical and Petrographical Collection of the Swiss Federal Institute of Technology (Zürich) owns a suite of contact-metamorphic rocks from the aureole around a granite strip belonging to the Lausitz Massif (Saxony), and lying to the south of Dresden on the left bank of the Elbe between Dohna and Zehista in the north and Berggiesshübel in the south. They belong to the greywacke formation of Culmic to Silurian age and are quite typical products of normal granite contact. With a view to clarifying some questions relating to the chemistry of metamorphic rocks, seven analyses were carried out at the writer's request by J. Jakob.

TABLE 1

	1	2	3	4	5	6	7
SiO_2	50.24	61.00	63.25	54.76	39.06	49.61	94.74
TiO_2	1.72	0.97	0.81	1.14	0.76	1.70	0.19
Al_2O_3	16.24	18.32	17.48	24.02	2.50	14.80	1.00
Fe_2O_3	3.49	1.95	1.26	1.86	3.47	2.13	0.09
FeO	7.83	4.53	4.27	4.93	—	9.61	1.12
MnO	0.09	0.03	0.06	0.06	0.06	0.15	0.00
MgO	6.00	2.72	2.06	1.98	1.39	7.74	0.00
CaO	2.74	0.86	1.38	0.49	27.70	8.85	0.27
Na_2O	4.10	3.02	3.23	2.10	1.63	4.01	1.71
K_2O	1.56	4.93	4.38	5.98	1.17	1.33	0.93
H_2O+	3.08	1.80	2.00	2.82	0.77	0.20	0.30
H_2O-	0.15	0.00	0.00	0.00	0.00	0.00	0.00
P_2O_5	0.57	0.04	0.02	0.00	0.00	0.00	0.00
CO_2	2.10	0.00	0.00	some C	21.85	—	0.00
	99.91	100.17	100.20	100.14	100.36	100.13	100.35

1. Slate. Burkhardtswalde near Weesenstein (matrix of clay minerals with some quartz and calcite with carbonaceous pigment.)
2. Nodular andalusite mica schist. Seidewitztal near Zuschendorf. (Mineral composition mainly quartz, biotite, andalusite, some muscovite and ore.)
3. Cordierite-hornfels. Lockwitztal, near the chocolate factory. (Cordierite, feldspar, quartz, biotite, occasional garnet and tourmaline.)
4. Cordierite mica-hornfels. Prenskerschacht in the Lockwitz valley near Kreischa. (Very rich in cordierite, some zoisite, biotite, muscovite and little quartz.)
5. Calcic greywacke. Friedrichswalde in the Bahra valley near Pirna. (Consists chiefly of quartz and calcite with zoisite, epidote, feldspars.)
6. Amphibolite (schistose). Bahraberg, between Zehista and Friedrichswalde, (consists chiefly of hornblende and plagioclase).
7. Quartzite with hornfels structure. Nenntmannsdörfer Mühle, Seidwitztal near Zehista. (95% quartz, beside some feldspar and biotite.)

TABLE 2. MOLECULAR VALUES (NIGGLI-VALUES)

	si	al	fm	c	alk	k	mg	ti	c/fm	co$_2$
1	141	27	51	8	14	0.20	0.49	3.6	0.16	8.0
2	225	40	34.5	3.5	22	0.52	0.43	2.7	0.1	—
3	250	40.5	30	6	23.5	0.47	0.40	2.4	0.20	—
4	189	49	29	2	20	0.65	0.35	3.0	0.07	—
5	103	4	12.5	77.5	6	0.32	0.44	1.5	6.2	7.8
6	112	19.5	48	21.5	11	0.18	0.54	2.9	0.45	—
7	2285	14	24.5	7	54.5	0.26	0.0	3.5	0.29	—

Nrs. 1–4 show a large excess of alumina and belong (Niggli 1934) to the alumosilicate rocks. They undoubtedly were derived from argillaceous rocks with, at most, a very slight addition of alkalis.

Nr. 5 is a calc-silicate rock derived from a lime-rich greywacke.

Nr. 6 has the chemical composition of a rather alkali-rich calc-alumosilicate rock and may have been derived from a shale, tuff or eruptive rock of gabbrodioritic composition.

Mr. 7 is a SiO_2 rock derived from a quartz-sandstone.

In what follows, only the alumo-silicate rocks will be considered together with some problems of chemical petrology which they raise. The marginal areas of the Lausitz granites often contain zones rich in inclusions and characterized by an alteration of the biotite granite to two-mica granite. Many modern hypotheses tend to ascribe an important rôle to metasomatic and chemical alterations. In view of this fact it seems important to compare the chemical composition of metamorphic sediments (para-rocks) with those of rocks whose composition does not seem likely to have suffered any considerable chemical change. Molecular (Niggli-) values are well adapted to this purpose as they render conspicuous the alumina excess (al-(alk+c)) typical of pelitic rocks.

Table 3 contains a number of examples selected at random from a large collection. It is evident that no great differences exist between these

TABLE 3

	si	al	fm	c	alk	k	mg	
clay, molasse	250	43	37.5	6	13.5	0.56	0.56	Horgen (Zürich), Switzerland
clay, molasse	250	43	35.5	4	17.5	0.63	0.47	Wettingen, Baden, Switzerland
green slate	206	43	31	6	20	0.28	0.35	Anglesey, England
roofing slate	240	43	38.5	1.5	17	0.63	0.36	Vermont, U.S.A.
Virginia slate	258	41.5	39.5	2	17	0.46	0.46	Mesabi, U.S.A.
slate rock	254	41	37	1	21	0.36	0.33	Anglesey, England
greywacke	300	42.5	30	5.5	22	0.38	0.47	Kamenz, Saxony

clays, shales and slates (pelitic sediments) and the rocks quoted above. A slight addition of alkalis seems likely, especially in the cases of nos. 2 and 3. Slight alterations such as these fall well within the scope of normal metamorphism and by no means justify considering these specimens as true metasomatic rocks.

The same molecular values may also be used to find metamorphic rocks which might be of similar mineral composition. For the SiO_2 value which has purposely been separated from the other molecular values remains practically unaffected by changes in mineral composition, except when the SiO_2 content is very low. Thus Table 4 contains a number of rocks whose compositions are closely related to those of rocks 1 to 4 in Table 2. It is immediately evident that the mineral composition can differ widely. Herein lies the problem of the relationship of metamorphic facies to temperature, pressure and water content. It has become usual nowadays to characterize the major units of the facies classification by special names instead of adhering to the historical method (Grubenmann-Becke) which used general definitions for this purpose.

It is not proposed to proceed on these lines here. For there seems to be little reason for including a staurolite-kyanite facies under the general concept of an amphibolite facies. It would surely be most inconvenient to assign a granite rock to the pyroxene gabbro facies, or to the sanidinite facies, and nobody would like to call a clay a limestone facies. We understand by rock facies a definite assemblage of minerals with a given structure and fabric. There is no justification in including in names such as green-schist facies, epidote-amphibolite facies, pyroxene hornfels facies, eclogite facies, etc., rocks with other minerals, structures and fabrics but formed under analogous physical conditions.

If it be desired to correlate facies of different mineral and chemical composition with respect to their conditions of formation, concepts such as Katafacies, Mesofacies and Epifacies with the necessary transitions are far more useful. Similar terms (in the main relating to temperature) have been applied in the treatment of ore deposits. They have long since (Grubenmann-Niggli 1924) been dissociated from any specific connection with depth or definite zones in the earth's crust as controlling factors. (The objections raised by Eskola and summarized by Turner (1948, p. 34) are less important than the objections given above.) The subfacies given in Table 5 which in principle all have the same chemical composition, can be arranged as shown in the tabulation at top of next page (some minerals may occasionally be absent).

In the case of an SiO_2 deficit the meso- to epifacies may also include emery subfacies with chloritoid, ores, diaspore, etc. (Oenay 1950).

In order to understand the connection between the various types, a

1. Cordierite-andalusite (sillimanite, mullite)-feldspars-pyroxene-subfacies (si-deficit: spinel, corundum)
2. Cordierite-sillimanite (andalusite)-feldspars-biotite-subfacies
3. Sillimanite (andalusite)-feldspars-biotite-subfacies
4. Sillimanite-garnet-feldspars-biotite-subfacies (± cordierite)
5. Garnet-biotite-feldspars-subfacies
6. Cordierite-feldspars-anthophyllite-biotite-subfacies
7. Andalusite (sillimanite)-biotite-muscovite-(feldspars)-subfacies
8. Garnet-biotite-muscovite (feldspars)-subfacies
9. Kyanite-staurolite-biotite-muscovite (feldspars)-subfacies
10. Biotite (chlorite)-muscovite (feldspars)-subfacies
11. Chloritoid (staurolite, biotite, andalusite, garnet)-sericite-subfacies
12. Chloritoid (chlorite)-sericite (albite)-subfacies
13. Sericite-albite-chlorite (glaucophane)-subfacies
14. Sericite-chlorite-subfacies

Katafacies
(kata conditions)
Kata-mesofacies
Mesofacies
(meso conditions)
Meso-epifacies
Epifacies
(epi conditions)

molecular base must be calculated and standard norms for kata-, meso- and epi conditions be deduced from it. As all compounds are given in formulae of a definite size, it is easy to express the connection between minerals and to calculate the modifications corresponding to any subfacies. The underlying principles have been explained elsewhere (Niggli 1936, Burri-Niggli 1945, Niggli 1948).

One example may here be treated more fully.

Rock Nr. 3, Table 1

	atomic numbers	Ru	Kp	Ne	Cal	Sp	Fo	Fa	Fs	Q	
Si	1053		93	104			2	30	8	816	
Al	342		93	104	50	95					
Fe'''	16								16		
Fe	60							61			
Mn	1										
Mg	51					47	4				
Ca	25				25						
Na	104			104							
K	93		93								
Ti	10	10									
	1755	10	279	312	75	142	6	91	24	816	1755
base (%)		0.6	15.9	17.8	4.3	8.1	0.3	5.2	1.4	46.4	100

To deduce the base from the kata standard norm, the following equations are required: (*see page 872*)

TABLE 4

	si	al	fm	c	alk	k	mg	
Plagioclase-cordierite-hornfels	203	36	42	8	14	0.67	0.41	Kolaas, near Oslo, Norway
Cordierite-hornfels	215	46	31.5	3.5	19	0.70	0.37	Mt. Ascutney, Vermont, U. S. A.
Andalusite-cordierite hornfels (leptite)	289	46	30.5	1	22.5	0.84	0.76	Långban, Sweden
Cordierite-andalusite-hornfels	286	46	28	3	23	0.91	0.31	Montana, U.S.A.
Cordierite-spinel-hornfels	131	45	42.5	4	8.5	0.44	0.34	Perthshire, Scotland
Cordierite-hornfels	229	40.5	47.5	0.5	4.5	0.84	0.50	Commando Neck, Vredefort, S. Africa
Gneiss with sillimanite, biotite, cordierite and garnet	107	41	54	1.5	3.5	0.68	0.40	Valpelline, Italy
Sillimanite-cordierite-hornfels	205	43	40	5.5	11.5	0.70	0.43	Mooifontein, Lydenburg, S. Africa
Sillimanite-gneiss	259	42.5	32	6.5	19	0.39	0.50	Langenbielau, Silesia
Andalusite-hornfels	143	42.5	43	1.5	13	0.68	0.34	Valdana, Elba
Andalusite-biotite-schist	197	47	29	3	21	0.51	0.28	Azegour, Marocco
Biotite-schist	319	37	40	5.5	17.5	0.30	0.45	Koltschak Isle, Sweden
Andalusite-mica-hornfels	227	45	32	4	19	0.38	0.36	Eichgrün, Auerbach, Saxony
Two-mica-gneiss	214	44.5	29.5	3	23	0.44	0.33	Simplon tunnel, Valais, Switzerland
Two-mica-schist	384	42	31	6.5	20.5	0.89	0.54	Lucomagno, Switzerland
Garnet-two-mica-schist	264	40	33.5	9	17.5	0.56	0.37	Stavanger, Norway
Garnet-gneiss	313	42	32.5	5.5	20	0.64	0.39	Nynäs, Sweden
Garnet-gneiss	257	38	36	10	16	0.20	0.39	Fort Ann, New York, U.S.A.
Garnet-mica-schist	230	41	35.5	7	16.5	0.60	0.32	Bru, Stavanger, Norway
Sillimanite-kinzigite	220	46.5	41.5	3.5	8.5	0.66	0.40	Valpelline, Italy
Staurolite-biotite-schist	154	35	45	8	12	0.52	0.44	Chittering Valley, W. Australia
Kyanite-mica-schist	400	43	38	1.5	17.5	0.65	0.89	Gastein, Austria
Sericite-albite-gneiss with glaucophane and chlorite	244	40	33	4	23	0.43	0.35	Val de Bagnes, Valais, Switzerland
Chlorite-sericite-albite-gneiss with glaucophane	231	40	38	4	18	0.65	0.37	Val de Bagnes, Valais, Switzerland
Chloritoid-schist	262	45	41.5	3.5	10	0.85	0.30	Reichenau, Grisons, Switzerland

2 Fs+1 Fa=2 Mt+1 Q
3 Kp+2 Q=5 Or
3 Ne+2 Q=5 Ab
3 Cal+2 Q=5 An
6 Sp+5 Q=11 Cord
3 Fo+1 Q=4 En
3 Fa+1 Q=4 Hy

The following kata standard norm (equivalent) is now obtained by calculation:

Ru	Or	Ab	An	Cord	Hy(+En)	Mt	Q
0.6	26.5	29.6	7.2	14.9	6.4	1.4	13.4

The rock could, therefore, be a cordierite hornfels with hypersthene and in this case the figures quoted would (although they are strictly speaking equivalent units) approximate closely to volume percentages. Now in reality the rock belongs to the cordierite-feldspars-biotite subfacies. This means that pyroxene and orthoclase are largely replaced by biotite+quartz. The equation which at once permits a calculation of the new mineral composition reads as follows (in terms of equivalent numbers):

10 Or+12 (En or Hy)+water=16 Bi+6 Q

in our case, therefore,

5.3 Or+6.4 Hy+water=8.5 Bi+3.2 Q

For rocks belonging to the cordierite-feldspars-biotite subfacies, we therefore have the following expression:

Ru	Or	Plag	Cord	Bi	Mt	Q
0.6	21.2	36.8	14.9	8.5	1.4	16.6

These figures give a close approximation of the observed mineral composition.

Other rocks of Table 4 contain andalusite in place of cordierite and much more mica (including muscovite). They, therefore, belong to the andalusite-biotite-muscovite subfacies. The following equations can now be used containing Sil in place of And:

10 Or+33 Cord+water=16 Bi+18 Sil+9 Q

or

40 Or+33 Cord+water=42 Ms+16 Bi+16 Q.

If we transform 14.9 parts of cordierite in our rock into biotite+andalusite, we obtain:

4.5 Or+14.9 Cord+water=7.2 Bi+8.1 Sil+4.1 Q.

The kata norm of an andalusite-biotite-feldspar subfacies thus reads as follows:

Ru	Or	Plag	Bi	Sil(And)	Mt	Q
0.6	16.7	36.8	15.7	8.1	1.4	20.7

Usually the rocks of Saxony contain some muscovite and often some zoisite and thus already have the character of a meso facies. Accordingly, the Ms-equation must be calculated. In our case it contains the values:

$$18.1 \text{ Or}+14.9 \text{ Cord}+\text{water}=19.0 \text{ Ms}+7.2 \text{ Bi}+6.8 \text{ Q}.$$

The meso norm for the rock in question in a biotite-muscovite-feldspars subfacies is, therefore, given by the following expression:

Ru	Or	Plag	Ms	Bi	Mt	Q
0.6	3.1	36.8	19.0	15.7	1.4	23.4

These figures show that a muscovite-biotite-plagioclase gneiss could have the same composition. Actually andalusite or cordierite are rarely absent in the above-mentioned contact metamorphic rocks. This enables the particular conditions of metamorphism (or of the subfacies) to be circumscribed more precisely.

The problem of the formation of garnet (mostly almandine) in place of cordierite has been discussed by Willemse (1936) and Masson (1938), who in their Zürich dissertations made valuable contributions to this question. Very often slight chemical differences (c-content, mg-proportion) favour the formation of one or the other subfacies. General equations could be written in the following way:

$$8 \text{ Hy}+11 \text{ Cord}=16 \text{ Alm}+3 \text{ Q}.$$
$$8 \text{ Fe-Bi}+6 \text{ Sil}+1 \text{ Q}=7 \text{ Ms}+8 \text{ Alm}$$

Of course if the composition of the garnet is known, it must be used for the calculation.

For the formation of staurolite one of the following equations may obtain:

$$8 \text{ Alm}+15 \text{ Sil}+\text{water}=21 \text{ Staur}+2 \text{ Q}$$
$$8 \text{ Fe-Bi}+21 \text{ Sil}+\text{water}=7 \text{ Ms}+21 \text{ Staur}+1 \text{ Q}$$

or

$$11 \text{ Fe-Cord}+2 \text{ Hm}+\text{water}=7 \text{ Staur}+3 \text{ Mt}+3 \text{ Q}.$$

From the base we obtain directly

$$3 \text{ Hz}+3 \text{ Sil}+1 \text{ Q}+\text{water}=7 \text{ Staur}.$$

Calculations of a meso standard norm (with staurolite in subfacies) and an epi standard norm (with chloritoid) have been given by C. Burri and P. Niggli (1945), p. 614 and E. Niggli 1944, p. 238. Not until such calculations have been made and the formation of variants discussed, can the fundamental principles of metamorphic facies be properly grasped. For only with their help can the possible variations in mineral assemblage be established that can occur in connection with a given chemical composition.

By way of comparison the kata standard norm of the three hornfelses nos. 2, 3 and 4 (table 1) may be quoted as follows:

No.	Ru	Or	Ab	An	Cord	Fe Cord	Sil	Hy	Mt	Q
2	0.7	29.8	27.8	4.1	19.6	—	—	6.4	2.0	9.4
3	0.6	26.5	29.6	7.2	14.9	—	—	6.4	1.4	13.4
									Hm	
4	0.8	36.4	19.5	2.7	15.4	21.8	2.0		1.4	—

The fact that no. 2 contains andalusite and mica, while no. 3 contains cordierite, does not emerge immediately from the very similar kata standard norms. Differences in the physical conditions of the metamorphism must be the reason and undoubtedly the course of the reaction had an important influence (no. 2 is a nodular schist, no. 3 a uniformly recrystallized hornfels). The high Cord-value of 4 in conjunction with the lack of Q in the kata standard norm favoured the formation of mica beside cordierite in this rock. The chemical composition would under other physical conditions certainly have lead to a rock containing garnet or staurolite or chloritoid.

To build up a meso norm, we can first calculate magnesium biotite (Bi), iron biotite (Fe-Bi) and kyanite (Sil) of the ideal composition. Oligoclase is possible.

MESOFACIES OF THE HORNFELSES

No.	Ru	Or	Ab	An	Ms	Bi	Fe-Bi	Sil	Mt	Q
2	0.7	0.5	27.8	4.1	25.2	10.3	7.8	—	2.0	21.6
3	0.6	3.2	29.6	7.2	18.9	7.7	8.0	—	1.4	23.4
4	0.8	—	19.75	2.7	36.4	7.6	8.8	6.75	2.0	15.2

Instead of Fe-Bi we often find garnet or staurolite. A garnet subfacies (almandine) of No. 4 gives:

No.	Ru	Ab	An	Ms	Bi	Alm	Sil	Mt	Q
4	0.8	19.75	2.7	44.1	7.6	8.8	0.15	2.0	14.1

Staurolite is calculated in the next subfacies:

No.	Ru	Ab	An	Ms	Bi	Fe-Bi	Staur	Sil	Mt	Q
4	0.8	19.75	2.7	38.65	7.6	6.2	6.75	—	2.0	15.55

Of course garnet and staurolite are in reality not free of magnesium and instead of oligoclase there may be albite (a part of the sodium is also in the muscovite) and zoisite+epidote. If the minerals are analyzed, the calculation of an equivalent mode is easy. But in connection with our problem the best way to elucidate the connection between kata- and meso-alumosilicate rocks is the calculation of a kata standard norm from some well known meso rocks of the Swiss Alps.

TABLE 5. SOME METAMORPHIC ROCKS OF THE SWISS ALPS

No. lit. 11	si	al	fm	c	alk	k	mg	ti	
IV 448	430	46	24	3	27	.63	.83	1.2	Biotite gneiss, M. Cauri, Ticino
IV 404	314	40	30.5	2.5	27	.58	.46	2.7	Two-mica-gneiss, Binnental, Valais
IV 447	305	37	39	2	22	.53	.89	2	Mica-gneiss (kyanite, sillimanite), M. Cauri, Ticino
II 305	170	38	40	5	17	.55	.50	4.2	Sericite-chlorite-albite gneiss, Maderanertal, Uri
IV 461	337	37.5	36.5	9	17	.47	.44	6.4	Biotite gneiss (garnet, sillimanite), S. Roveredo, Ticino
IV 460	223	39.5	41	3.5	16	.66	.39	3.8	Muscovite-chlorite-schist, S. Iorio, Ticino
III 172	196	44	38	3.5	14.5	.78	.48	3.1	Kyanite-two-mica-schist, Frodalera, Ticino
IV 145	300	48.5	34	5	12.5	.50	.43	4.2	Garnet-two-mica-schist (kyanite), Frodalera, Ticino
IV 221	148	53.5	34	4	8.5	.76	.29	3.9	Garnet-two-mica-schist (staurolite, kyanite) P. Molare, Ticino
IV 146	86	42	49.5	5	3.5	.27	.15	1.6	Staurolite-mica-schist (garnet, kyanite), P. Molare, Ticino
III 174	168	57	26.5	4	12.5	.42	.25	4.9	Staurolite-sericite-schist, Aquacalda, Ticino
IV 149	91	59.5	27.5	1.5	11.5	.64	.18	2.5	Staurolite-schist (kyanite), Pizzo Molare, Ticino

Table 5 gives the molecular values, the rock name, the percentages and number in lit. 11. In lit. 11 the analyses are given in weight %.

The kata standard norm is calculated in Table 6. In some cases the

TABLE 6. THE KATA STANDARD NORM OF THE ROCKS OF TABLE 5

	Or	Ab	An	Cord	Fe-Cord	Hy	En	Sil	Mt Hm	Ru	Q	Not satu-rated −Q
IV 448	28.7	16.3	2.5	14.3	—	.9	1.3	—	.2	.2	35.6	—
III 404	32.4	23.3	2.8	11.7	—	2.8	1.5	—	2.0	.6	22.9	—
IV 447	25.3	22.5	2.2	15.2	—	1.5	9.2	—	.1	.4	23.6	—
II 305	27.3	23.7	7.7	27.2	—	8.3	2.0	—	2.1	1.3	10.4	—
IV 461	16.2	18.2	9.2	12.7	—	7.1	1.7	—	.6	1.3	33.0	—
IV 460	27.2	14.0	4.7	22.6	6.6	4.7	—	—	3.1	1.0	16.1	—
III 172	31.5	9.5	4.7	27.9	7.0	—	—	2.7	4.3*	.9	11.5	—
IV 145	14.0	13.5	5.2	17.2	12.5	—	—	3.8	1.9*	.9	31.0	—
IV 221	20.5	6.5	6.7	17.1	28.1	—	—	14.3	2.7*	1.2	2.9	—
IV 146	4.3	11.3	10.5	17.1	61.4	—	—	1.4	5.7*	.8	—	−12.5
III 174	15.3	20.8	6.0	10.8	7.0	—	—	25.8	4.5*	1.4	8.4	—
IV 147	28.0	16.0	2.8	10.3	12.3	—	—	40.5	6.2*	1.0	—	−17.1

SiO_2-content was too low to form the normal standard minerals; in these cases the deficit of Q is given, instead of the calculated amount of olivine or spinels.

The first group of 7 analyses (without much Fe-cordierite or sillimanite in the kata standard norm) corresponds to rocks with only subordinate quantities of special minerals. If the content of Fe-cordierite and sillimanite is a little higher, garnet (almandine) is more easily formed in the mesofacies (the next 2 analyses). Higher content of Fe-cordierite and (or) sillimanite is found in the last 3 analyses which correspond to meso rocks rich in staurolite.

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FAYALITE AT ROCKPORT, MASSACHUSETTS*

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Abstract

The facts are collected concerning the place and mode of occurrence of fayalite at four localities on Cape Ann, Massachusetts, with a description of one of them which yielded the largest and best-developed crystals. All of these occurrences are in pegmatite of a granitic nature but evidence is presented which seems to show that all are xenoliths, genetically related to a fayalite-bearing nordmarkite granite, older than and intruded by the dominant Rockport granite.

Fayalite is not a mineral of common occurrence. It is true that it has been found as a constituent of certain granites and is well known to be abundantly formed in the slags of iron smelters. But individual crystals of natural occurrence are very rare. Boulders of an iron silicate found on a beach on the island of Fayal in the Azores were supposed to have come from the local volcanic rocks and the name *fayalite* was based on their description. However, further study showed that in all probability the boulders were in reality lumps of slag, carried as ballast in some ship and left ashore there. Nevertheless the name has persisted. Minute crystals of great beauty have been found in the cavities of rhyolites and from them the crystal form has been well established.

These facts are stated in order to bring out the very unusual character of the four occurrences of fayalite known in the granite of Rockport on Cape Ann, Massachusetts, to the description of which the following pages are devoted.

The first to find fayalite at Rockport was Mr. J. H. Sears (1905), the Curator of the Peabody Museum at Salem. In 1890 he sent specimens to Professor Penfield at Yale University, reporting that they had been found in the granite of the large Rockport Granite Quarry at a depth of sixty feet. They were parts of a lenticular shell of varying thickness, from twelve to sixteen inches in diameter, filled on the inside with loose earthy material and enveloped by a layer of magnetite about one inch thick. This was in the granite but adjacent to a pegmatite boss or vein. There is a piece of this find in the Harvard Collection which shows a section of a large crystal like those to be described later. On fresh fracture the color is greenish black with resinous luster and two cleavages are shown. Penfield (1896) showed that these cleavages were very perfect parallel to the basal pinacoid; less perfect parallel to the branchypinacoid, thus correcting the previous description of this property. He made an analysis, finding it to be a very pure silicate of iron; and he determined for the first time its optical constants. Much later (Bowen and Schairer, 1932) it was found that the melting temperature of this fayalite was 1205° ± 2°C, exactly the same, within the limits of error, as that of pure synthetic ferrous silicate.

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* Contribution from the Department of Mineralogy and Petrography, Harvard University, No. 319.

Fayalite was next reported from Rockport by Dr. Charles H. Warren, (1903) then of the Massachusetts Institute of Technology, who described a second discovery in the same quarry. This mass was found by two of his students and he did not see it in place. It was large, some 250 pounds having been obtained, and, as this was sold to dealers, the mineral was widely distributed in collections. The specimens were sections of large crystals, tapering from a thickness of ten inches down to a thin edge. There was a small amount of black lepidomelane mica on the surfaces of the crystals as well as minute zircon crystals. The main purpose of this paper was to describe a reaction zone developed where the fayalite was in contact with quartz; the zone consisted of a thin layer of a finely fibrous ferroan anthophyllite mixed with granular magnetite. It is now known (Bowen and Schairer, 1935) that this amphibole is gruenerite, it being monoclinic rather than orthorhombic as Warren thought.

We next hear of fayalite at Rockport in 1908 and, as the writer had a hand in obtaining these specimens which are undoubtedly the finest found anywhere of this mineral and have never been adequately described, the facts may well be recorded in detail.

In 1908 Dr. Dale of the United States Geological Survey wrote on the Commercial Granites of Massachusetts. In describing the Rockport granite he states that in the Babson Farm Quarry, which had been opened not long before on the extreme northeast point of Cape Ann, there had been found a "knot" or pegmatite of unusual character which contained, among other more usual minerals, "a mass six by two inches of yellowish brown color, determined by Mr. Johannsen as bronzite." As this mineral seemed wholly out of place in the paragenesis of granite pegmatite I at once suspected that another find of fayalite had come to light and took an early occasion to visit the quarry with a party of students.

We found the quarry itself well worth the journey. It had been opened by a contractor to supply rectangular granite blocks weighing thirty or forty tons each for the construction of a subsurface breakwater to protect the harbor of Rockport. The operation was on a gigantic scale, and the great blocks in process of being broken out of the quarry with wedges or hoisted out onto the surface for transfer by rail and scow were very impressive. The granite was coarse and wonderfully homogeneous with almost no sign of the dikes or pegmatite segregations so common in other Cape Ann quarries. By the side of the railway track stood the block which Dale had described, condemned because of its poor quality and awaiting removal to the dump of waste rock. It measured about eight feet by four by two and a half, and was composed in large part of coarse pegmatite in sharp contact with the normal granite. The striking feature of this contact was its cross-cutting nature; the structure of the pegmatite bore no relation to it. The evidence was conclusive that the pegmatite was but a part of a larger mass which had been disrupted by the intrusion of the granite and that this angular block was one of the fragments, floated off as a xenolith in the granite. Search of the large waste dump and examination of the vast exposed faces of the quarry walls revealed no other like material. So this lone block, left by chance after being brought to the surface, to be seen by chance by a visiting geologist on a timely visit, was all there was found of this pegmatite.

The "bronzite," or rather fayalite as it proved to be, showed on two surfaces of the quarry block but getting it out looked like a hopeless task. However, two of the party, George M. Flint and A. C. Gosse, returned the next day armed with drills, hammers and plug-and-feather wedges. They obtained permission to break into the discarded stone, and ultimately brought back to Cambridge the larger part, perhaps a hundred pounds, of the pegmatite inclusion.

The chief constituent and latest mineral to form was quartz in large anhedra, glassy but of a peculiar hazy opalescent blue color; masses of several pounds weight were obtained. The feldspar was white microcline characterized by an unusual curvature both of platy masses and of individual crystals where it was embedded in quartz. In the feldspar were warped and cracked plates of ilmenite up to five inches in diameter but less than an eighth of an inch in thickness. There were also quite large prismatic crystals of a green pyroxene which proved to be hedenbergite. And lastly there was the fayalite, the large crystals of which were segregated in a single mass, lying against one another like sardines packed in a can. The principal specimen on exhibition in the Mineralogical Museum at Harvard University is estimated to weigh about sixty pounds of which perhaps twenty is due to fayalite. It is shown in the sketch, Fig. 1.

The most complete crystal broke away from the matrix but may be fitted back into its place and then appears to be about one-quarter of the whole individual. It measures five and one-half inches in the direction of the *b* axis, six inches parallel to *c* and one inch and a half parallel to the axis *a*. Its broken edges are cleavages; the faces are in part quite plane but the pinacoid curves into a vaguely developed prism form; the faces could be accurately measured with the contact goniometer yielding definitely the form shown in the drawing, Fig. 2. A second crystal, less complete, was thicker and probably larger in all its dimensinos. A fragment of a third crystal separates the two larger individuals. Embedded in the

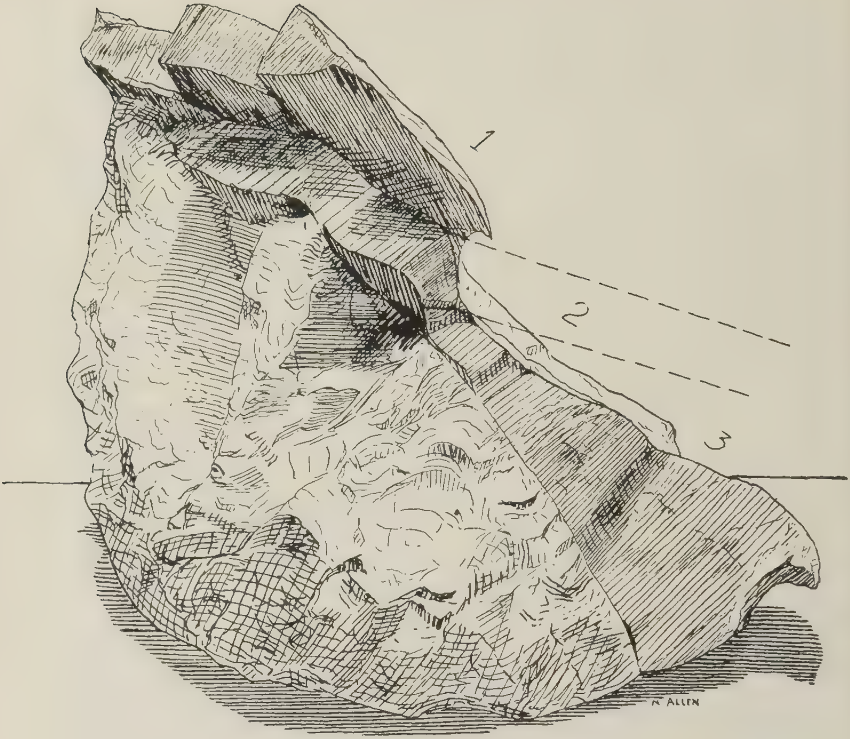

FIG. 1. Fayalite crystals in microcline. The three crystals are indicated. The termination of the uppermost complete crystal is not visible.

surface of the fayalite were a few much flattened grains of a black amphibole, not fibrous and of high index, probably hastingsite.

The latest description of fayalite in the Rockport area was by Warren and McKinstry (1924). They described the pegmatites and their minerals, and in several places refer to the last-described mass of pegmatite

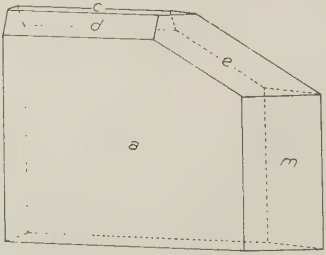

FIG. 2. Crystal of fayalite. *c*{001}, *a*{100}, *m*{110}, *d*{101}, *e*{111}. The prism is not plane on the crystal but curves into the pinacoid.

which was placed at their disposition. They were convinced, as I had been, that this mass was a xenolith in the Rockport granite. They also described a fourth occurrence of fayalite found in place by Dr. Warren on Brier Neck near Gloucester, on the opposite side of the Cape from the Babson Farm Quarry. This was in a dike-like mass of pegmatite which however is not well exposed and can not be proved to be an inclusion although they believe it to be one. This fayalite is in large part altered by a reaction with feldspathic material to a mixture of biotite, gruenerite and magnetite.

The fayalite from all four of these occurrences has the same optical and physical properties. Each was found in an isolated mass of considerable size and in the case of the Babson Farm Quarry the evidence for the xenolithic nature of the mass was clear. It is perhaps safe to conclude that all were of the same nature. If this be true then the formation of fayalite on Cape Ann belongs to a period of granitic intrusion and crystallization preceding that of the present dominant regional rock, the Rockport granite. Such a rock is described by Warren and McKinstry under the name of nordmarkitic granite. This granite, which is cut by the dominant granite, contains fayalite as one of its lesser constituents, which is in harmony with the preceding conclusion as to fayalite pegmatite. No fayalite has ever, so far as the record shows, been found in the very numerous pegmatites characteristic of the latest Cape Ann granite.

With the decay of the quarrying industry on the Cape, most of the quarries have been abandoned and it does not seem likely that any further finds of these scattered blocks containing fayalite will be discovered.

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REMARKS ON CRYSTALLOGRAPHIC NOMENCLATURE

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Abstract

In special cases the lattice (not structure) of a crystal in any system may be indistinguishable from the lattice typical of any higher system. Thus it is formally better to define and name the systems on the basis of symmetry (as groups of classes); for this purpose a set of self-explanatory names and symbols is proposed.

With the great increase of interest in crystallography which has developed since the first world war and the enormous extension of the field beyond its former boundaries, there had been a movement to revise the nomenclature of the science, to clarify and extend it where necessary, and to devise notations suitable for international use. In the largest single department of crystallographic terminology which has come under consideration, namely that which refers to the symmetry groups required in structural crystallography, the need for a good notation has been well met by the system of Hermann-Mauguin. Made generally available in the original *International Tables* (1933) this system has been well tried and generally adopted, and we may expect to find any necessary refinements and corrections to the scheme in the new edition now in preparation.

Less attention has been paid to the nomenclature of morphological crystallography, and it cannot be said that there is a generally accepted terminology for crystal systems, crystal classes, and crystal forms which has the consistency and self-explanatory character of the international notation for point groups and space groups. Those interested mainly in structural crystallography have naturally not been much concerned with the nomenclature of morphology; on the other hand those who have been principally occupied with crystal form have been well aware of certain imperfections in the relevant terminology, which constantly crop up in teaching and research, but they have seemed reluctant to give up established terms and usages which have served their turn despite inherent illogic and inelegance. But for those who see modern crystallography as a logical development of the classical science, the need for some improvements in the nomenclature of the older subject is apparent to bring it in line with the excellent terminology of the newer branch. It is with this in mind that I venture some discussion and proposals regarding the definition and terminology of the crystal systems, crystal classes, and crystal forms.

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Crystal systems

The crystal systems are the broad divisions into which the working crystallographer using the classical methods (reflecting goniometer and polarising microscope) can always place a well-formed crystal. Frequently a morphological study is limited to the determination of the crystal system, the geometrical elements, and a description of the forms; in a structural study the crystal system and lattice dimensions are also first determined, but only as a first step towards finding the space group and the structure. Thus the classical systems have become groups of convenience and their definitions and names have been retained without much critical concern.

Historically, the orthogonal and hexagonal systems were recognized first, and the inclined systems were admitted later, with the suggestion that they are in some ways defective. Without exception the textbooks presented the systems in the historical order, cubic to triclinic; and I can well recall the twinge of sympathy I felt as a student for the triclinic crystal, shown like a scarecrow with axes all awry, bereft of all symmetry except a centre of inversion, and pictured by a wall-eyed stereogram in unhappy contrast to the full fat smiling visage of the cubic holohedry. This historical order of presenting the crystal system, from the most specialized to the most general, is the reverse of universal mathematical procedure, and it has caused much of the unsatisfactory terminology with which we are concerned.

The classical crystal systems, with some alternative names and the usual definitions in terms of lattice dimensions, are as follows:

Cubic (isometric): $a(= b = c)$; $(\alpha = \beta = \gamma = 90°)$.

Hexagonal: $a(= b) \neq c$; $(\alpha = \beta = 90°; \gamma = 120°)$.

Tetragonal (quadratic): $a(= b) \neq c$; $(\alpha = \beta = \gamma = 90°)$.

Orthorhombic (rhombic, rectangular): $a \neq b \neq c$; $(\alpha = \beta = \gamma = 90°)$.

Monoclinic (monosymmetric): $a \neq b \neq c$; $(\gamma = \alpha = 90°)$; $\beta > 90°$.

Triclinic (anorthic): $a \neq b \neq c$; $\alpha \neq \beta \neq \gamma \neq 90°$.

The usual definitions of the crystal systems rest on the metrical properties of the axes or lattice elements (lengths and angles) and, apart from the names hexagonal and tetragonal, which involve symmetry, the commonest system names express essential metrical properties. Thus, cubic (or isometric), monoclinic, and triclinic are entirely suitable names for metrical systems. Some name other than hexagonal would be useful to denote the metrical system based on the two modes of stacking hexagonal nets, but none suggests itself. Rhombohedral is not mentioned here since the rhombohedral lattice is merely a mode of the hexagonal lattice, from the purely metrical point of view. Quadratic is preferable to tetragonal as a metrical system name; and if rectangular could be accepted in place of orthorhombic, to avoid the now inappropriate allusion to the base-centred mode of the lattice, a fairly satisfactory nomenclature for the metrical systems would result. This would read: cubic (isometric and optically isotropic); hexagonal and quadratic (dimetric and optically uniaxial); rectangular, monoclinic, triclinic (trimetric and optically biaxial).

However, there have long been misgivings regarding the validity of the metrical systems. Metrical pseudosymmetry (near equality of typically unequal lattice lengths and near approach to 90° or 120° of typically non-rectangular or non-hexagonal lattice angles) is a commonplace, and practising crystallographers are all familiar with cases where such inequalities, if real, are within the limits of goniometric accuracy. Thus we may have a crystal whose form development is typically "monoclinic" while the angle β cannot be distinguished from 90° (datolite); or again a crystal whose structure and optics are hexagonal (rhombohedral), but with a lattice which is exactly "cubic" (shandite). To meet such examples the definitions of the metrical systems are sometimes qualified by the proviso that the typical relations must obtain at all temperatures within the stability range of the phase. At first glance this seems to safeguard the metrical definitions, at least theoretically; but apart from the practical difficulty of making exact measurements at high and low temperatures, the temperature qualification is not adequate in the inclined systems. With change of temperature the β angle of a monoclinic crystal might change from 91° to exactly a right angle; by the metrical definitions the crystal would belong neither to the monoclinic nor to the rectangular system.

The truth is that, despite the names and definitions of the metrical systems, enlightened observers have always assigned crystals to their systems on the basis of symmetry, particularly the highest symmetry common to all the observable properties of the crystal. Some recognition of this fact is seen in the names monosymmetric (for monoclinic), tetragonal, and hexagonal. Would it not be reasonable to recognize clearly that the crystal systems are actually symmetry systems, and to devise a practical nomenclature which expresses this fact?

This has been attempted in the following table which shows the proposed names and symbols for the symmetry systems in relation to the most widely accepted names for the symmetry classes and their Hermann-Mauguin symbols. Below the names of the symmetry systems are given the names of the normally corresponding lattices; these are identical with the metrical system names, except that the rhombohedral lattice is recognized as distinct from the hexagonal lattice. It is emphasized that in any symmetry system the metrical relations or lattice may in special cases be indistinguishable from those typical of any higher symmetry system. As implied in the table the symmetry systems are defined simply as groups of crystal classes which are also grouped as usual into the centrosymmetrical Laue symmetries. The new names are in line with the current names, trigonal, tetragonal, and hexagonal, and they are in keeping with the principles of the international notation for the crystal classes. The symbols for the symmetry systems are the appropriate numerals, written in parentheses to distinguish them from the similar class symbols. A word of explanation will suffice for each of the symmetry systems.

Monogonal (1) is proposed for the system which comprises the classes characterized by a 1-fold rotation axis (no symmetry) or a 1-fold rotation-inversion axis (centre of symmetry).

Digonal (2) describes the system which includes the classes with a single 2-fold rotation axis or a single 2-fold rotation-inversion axis (mirror plane).

Tri-digonal (222) refers to the system of classes with three non-equivalent 2-fold axes, two of which may be 2-fold rotation-inversion axes (mirror planes).

Trigonal (3) is already used for the system of classes with a single 3-fold rotation axis or a single 3-fold rotation-inversion axis, in which the lattice may be rhombohedral or hexagonal.

Tetragonal (4) is likewise used for the system of classes characterized by a single 4-fold rotation axis or 4-fold rotation-inversion axis.

Hexagonal (6) is also used in the restricted sense of the system of classes with a 6-fold rotation axis or a 6-fold rotation-inversion axis, in which the lattice is always hexagonal.

Tetra-trigonal 4(3) is finally the logical name for the system of classes which have four equivalent 3-fold rotation axes or 3-fold rotation-inversion axes.

Triclinic, monoclinic, orthorhombic (or rectangular), tetragonal (or quadratic), hexagonal, rhombohedral, and cubic are of course still suitable names for the simple lattices; and in developing crystal morphology from the lattice, it is convenient also to use these same names for lattice systems, which are identical with the metrical systems except for hexagonal and rhombohedral, which denote distinct lattice systems but only one metrical system. If, in formal treatment, it is then shown that

Symmetry Systems		Symmetry Classes	
Monogonal	**(1)**	Pedial	1
(Triclinic)		Pinakoidal	$\bar{1}$
Digonal	**(2)**	Domatic	$m\ (=\bar{2})$
(Monoclinic)		Sphenoidal	2
		Rhombic prismatic	$2/m$
Tri-digonal	**(222)**	Rhombic pyramidal	$mm2\ (=\bar{2}\bar{2}2)$
(Rectangular)		Rhombic disphenoidal	222
		Rhombic dipyramidal	$2/m\ 2/m\ 2/m$
Trigonal	**(3)**	Trigonal pyramidal	3
(Rhombohedral or hexagonal)		Rhombohedral	$\bar{3}$
		Ditrigonal pyramidal	$3m$
		Trigonal trapezohedral	32
		Ditrigonal scalenohedral	$\bar{3}\ 2/m$
Tetragonal	**(4)**	Tetragonal disphenoidal	$\bar{4}$
(Quadratic)		Tetragonal pyramidal	4
		Tetragonal dipyramidal	$4/m$
		Tetragonal scalenohedral	$\bar{4}2m$
		Ditetragonal pyramidal	$4mm$
		Tetragonal trapezohedral	422
		Ditetragonal dipyramidal	$4/m\ 2/m\ 2/m$
Hexagonal	**(6)**	Trigonal dipyramidal	$\bar{6}$
(Hexagonal)		Hexagonal pyramidal	6
		Hexagonal dipyramidal	$6/m$
		Ditrigonal dipyramidal	$\bar{6}m2$
		Dihexagonal pyramidal	$6mm$
		Hexagonal trapezohedral	622
		Dihexagonal dipyramidal	$6/m\ 2/m\ 2/m$
Tetra-trigonal	**4(3)**	Tetartoidal	23
(Cubic)		Diploidal	$2/m\ \bar{3}$
		Hextetrahedral	$\bar{4}3m$
		Gyroidal	432
		Hexoctahedral	$4/m\ \bar{3}\ 2/m$

the lattice systems or the metrical systems are better replaced by symmetry systems with suitable names, the desired clarification of the nomenclature for the crystal systems will have been achieved.

Crystal classes

These are mentioned here only as a preparation for the subsequent remarks on the naming of crystal forms. The structural crystallographer has little need for class names, since the determination of the point group is but a second step toward the determination of the space group, and he commonly uses the class symbol which often serves also in morphological crystallography. However the determination of the crystal class from external form may be the main objective of a morphological study and consequently a set of appropriate names is naturally desirable.

Only two distinctly different principles have been employed in naming the classes: (1) the principle of subdividing the systems into holohedral or holosymmetric classes and merohedral or merosymmetric classes; (2) the principle of naming the classes after their general forms. The second of these principles is now widely accepted as the most useful, since it directly recalls the morphological characteristics of the classes, and the names given in the table are also in general use. In adopting this nomenclature we must of course adhere to the definitions of the forms after which the classes are named. These definitions are well known and need not all be repeated here.

Crystal forms

Finally I venture to comment on some terms which are still widely used to describe certain forms in the trimetric systems, even though the usage is in sharp conflict with the formal nomenclature on which the class names are based. The form names involved in this conflicting usage are principally pinakoid, dome, prism, and pyramid. In the present day nomenclature these and other forms or types of forms are defined as geometrical figures without reference to their attitude in space. A pinakoid (πίναξ, a board) is a pair of equivalent parallel planes; a dome (*domus*, a house [roof]) is a pair of non-parallel planes symmetrical to a mirror plane; a prism is a closed set of equivalent planes parallel to a common axis; and a pyramid is a closed set of equivalent planes equally inclined to a common axis. In the older usage these same terms have somewhat different meanings and certain restrictions of attitude are implied. Thus pinakoid is restricted to the pairs of axial planes, $\{100\}$, $\{010\}$, $\{001\}$; dome is used for any form of the type $\{0kl\}$ or $\{h0l\}$; prism is used for any form of the type $\{hk0\}$ and prism zone means vertical zone; and pyramid is used for any form of the type $\{hkl\}$. In the orthorhombic system (the old system names are suitable here) this gives {100}, front- or macropinakoid (parallel to the conventionally longer horizontal axis; {010}, side- or brachypinakoid (parallel to the conventionally shorter horizontal axis); {001}, basal pinakoid; {0*kl*}, brachydome [rhombic prism]; {*h*0*l*}, macrodome [rhombic prism]; {*hk*0}, prism [vertical rhombic prism]; {*hkl*}, pyramid [rhombic dipyramid]. Even stronger divergence of meaning results when the older usage is adapted from the orthorhombic system to the more general monoclinic and triclinic systems. In the monoclinic the older nomenclature gives {100}, front- or orthopinakoid (parallel to the ortho-axis [010]); {010}, side- or clinopinakoid (parallel to the clino-axis [100]); {001}, basal pinakoid; {0*kl*}, clinodome [rhombic prism]; {*h*0*l*}, orthodome or hemi-orthodome [pinakoid]; {*hk*0}, prism [rhombic prism]; {*hkl*} pyramid or hemi-pyramid [rhombic prism]. In the triclinic the older nomenclature gets into still deeper trouble. Again {100}, {010}, {001}, are the only forms called pinakoids, although all forms in the holohedral class are pinakoids. {0*kl*} and {*h*0*l*} are called hemidomes, {*hk*0} are hemiprisms, and {*hkl*} would become tetarto-pyramids, if the scheme were carried through.

This discordant nomenclature has long been current in crystallographic laboratories and in descriptive morphological writing and it has been given a new lease of life in F. C. Phillips' *Crystallography* (1946), where it appears with careful apologies in a treatment which substantially accepts the modern nomenclature. Actually the older nomenclature cannot be reconciled with the newer system, nor can it be developed into a consistent scheme of form names and class names which could replace the present system. The undesirable usage can be avoided in two ways. One can describe crystal forms by their strict names followed by the appropriate indices {*hkl*} to show the attitude of the form, when desired. This is the simplest and probably the best solution. Another way would be to devise a consistent and self-explanatory scheme of qualifiers as prefixes or affixes to the strict form names to denote the attitude. Such a scheme would best be worked out in consultation with a number of experienced morphologists.

Acknowledgment. Probably none of the proposals made in the foregoing notes is wholly new in principle, but for the sake of brevity they are offered here, as they were written, with almost no reference to authorities. I wish, however, specifically to acknowledge the incentive to discuss the crystal systems which I received from a preliminary unpublished table of symmetry devised by C. D. West (Cambridge, Massachusetts). In this table Dr. West uses "monogonal" and "digonal" as system names, but with meanings different from those given in this paper.

STRUCTURE AND MINERALOGY OF THE GOLCONDA PEGMATITE, MINAS GERAIS, BRAZIL*

W. T. PECORA,[1] GEORGE SWITZER,[2] A. L. BARBOSA,[3]
AND A. T. MYERS[1]

ABSTRACT

The granite pegmatite at the Golconda mine in Minas Gerais, Brazil, is a zoned, sheetlike mass in schist. Most of the estimated 1,000 metric tons of mine-crude muscovite mined since 1908 has been recovered from a mica shoot in the upper border zone. Detailed descriptions are given for garnet, cookeite, beryl, cassiterite, manganotantalite, and microlite—all of which minerals occur in the central zone.

INTRODUCTION AND ACKNOWLEDGMENTS

The Golconda pegmatite mine, one of the oldest mica producers in Brazil, was examined and mapped in detail in 1943–45 by Pecora and Barbosa in conjunction with a general program of the United States and Brazilian Governments aimed at increasing mica production and mica reserves in Brazil during World War II. The mineral collections made at that time were recently studied by Pecora, Switzer, and Myers in the laboratories of the U. S. Geological Survey and the U. S. National Museum.

The authors are grateful to a number of persons for aid during the field and laboratory investigations. Mr. José Nogueira, of Santos-Nogueira Company, operators of the Golconda mine, supplied production data; C. D. Foster and J. T. Cook, engineers of the U. S. Commercial Company, extended many courtesies in the field. The authors are grateful to Michael Fleischer, Earl Ingerson and K. J. Murata for critical reading of the manuscript and to S. B. Levin for calculation of the components of a garnet specimen.

LOCATION AND HISTORY

The Golconda pegmatite mine is located in the Rio Doce Valley, Brazil, 34 km by road northwest from the town of Governador Valadares (formerly Figueira)—a well-known mica center in eastern Minas Gerais that has rail, road, and airplane routes to Rio de Janeiro, Victoria, and Belo Horizonte.

Since the discovery of the Golconda pegmatite in 1908, the property has yielded muscovite mica, gem tourmaline, and gem beryl having an estimated total value of $100,000. In 1944 about 400 tons of mine-crude mica were recovered from open-pit operations (Fig. 1) by use of heavy, mechanical earth-moving equipment. In 1945 new underground workings were driven in the mica shoot along the hanging wall of the pegmatite.

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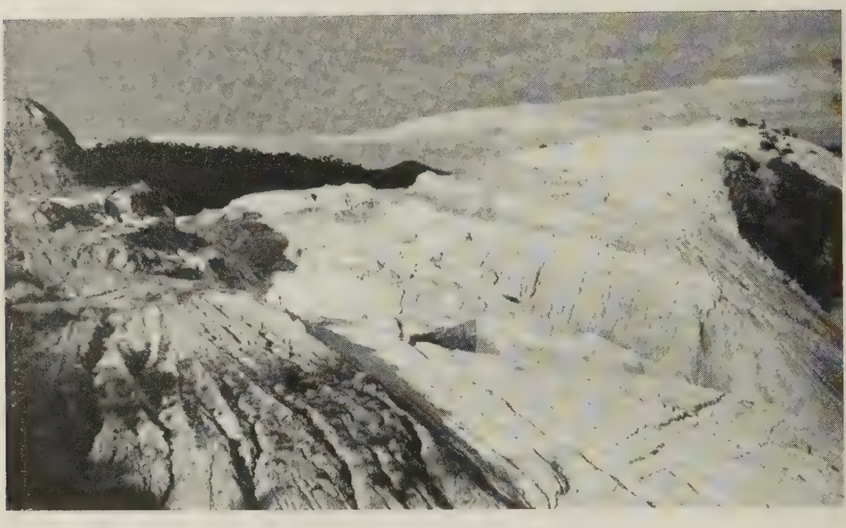

FIG. 1. View of the Golconda open pit in 1944 showing the area stripped by heavy mechanical equipment.

REGIONAL GEOLOGY

The region near Governador Valadares is underlain essentially by metamorphic rocks into which have been emplaced a great number of granitic pegmatites. The metamorphic rocks, principally schist and gneiss are considered by Brazilian geologists to be Archean in age. Although the age of the pegmatites is not known with certainty, a late pre-Cambrian or early Paleozoic age is in general accord with geologic evidence and age determinations based on analyses of radioactive minerals. The regional relations of the mica-bearing pegmatites in Minas Gerais are discussed by Pecora and others (1950).

DESCRIPTION OF THE PEGMATITE

Form, Size, and Structure

The Golconda pegmatite, one of several sheetlike pegmatites in the Rio Doce Valley, ranges in thickness from 3 to 11 meters and is nearly horizontal (Fig. 2). In the western part, the pegmatite has a well-defined arch or "roll"—a characteristic feature of many sheetlike pegmatites in this region. The crest of the roll pitches about 10°N. and has been exposed by mining for about 100 meters. In an east-west direction the pegmatite is exposed for about 300 meters.

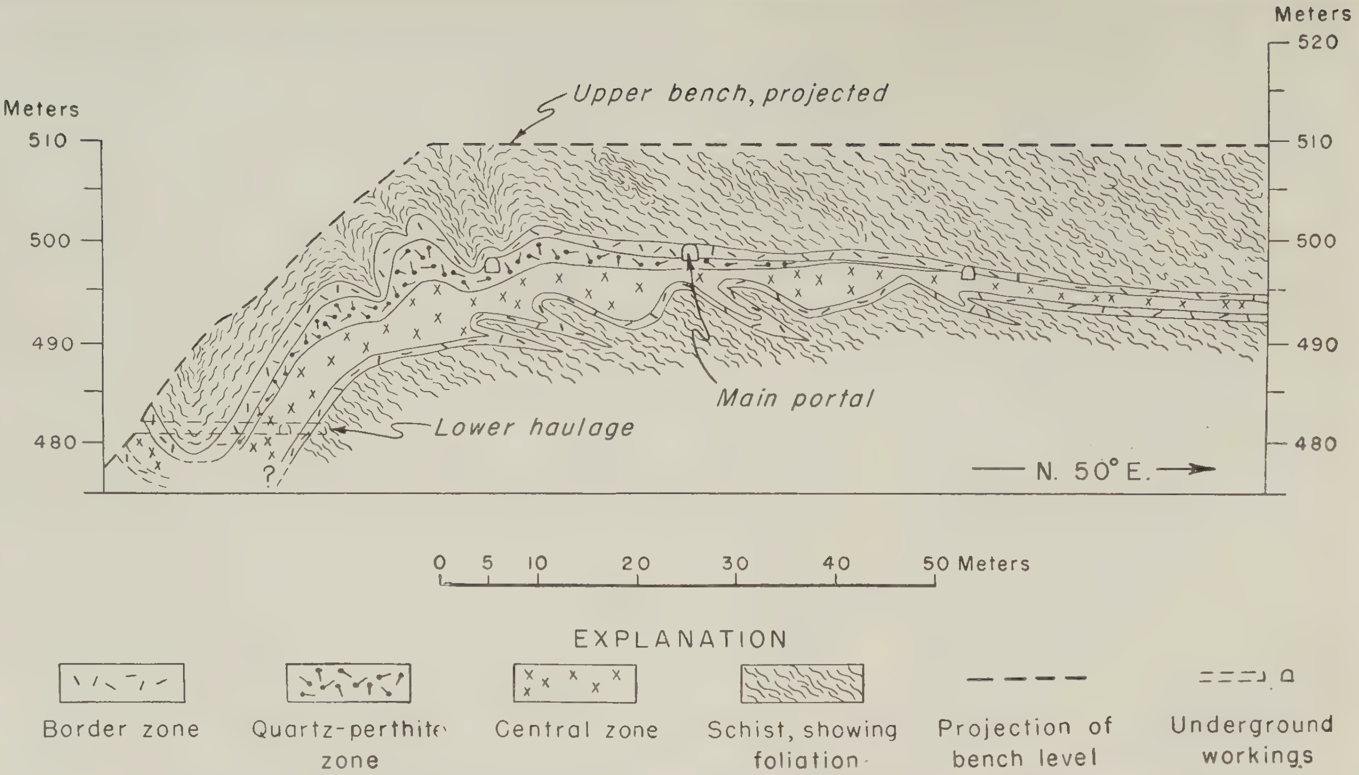

FIG. 2. Generalized cross section of the Golconda pegmatite, in part projected and reconstructed, showing internal zoning and the structural relation to schist.

The country rock containing the pegmatite is mica schist. In the western part of the open pit the schist is rich in hornblende, and in the eastern part it is essentially a quartz-biotite-garnet schist. The hanging-wall contact of the pegmatite locally conforms to the schistosity of the roof rock, but the footwall contact is clearly discordant to the schistosity of the floor rock. The schist above the pegmatite is intimately folded into a series of asymmetrical drag folds whose axial planes strike about N. 25°W. and dips northeast. The roof block of schist, now evident as a small anticlinorium overturned to the west, was probably sheared from the floor block, by deformational stresses, thus providing a locus for implacement of the granitic fluid. The roll of the pegmatite is believed to be an original implacement form and not one caused by later deformation. The shearing of the country rock and the subsequent introduction of granitic fluid were probably related to the same general diastrophism.

Zoning

The grain size of the minerals in the pegmatite increases inward from both its upper and its lower contact and is coarsest where the pegmatite is thickest. On the basis of differences in texture and mineral composition this pegmatite can be divided into several internal units.

In the eastern, thinner, part of the pegmatite sheet and in the larger appendages extending into the floor rock a central zone separates an upper and lower border zone. The central zone is composed essentially of quartz and weathered potash feldspar with accessory white mica, green tourmaline, and blue-green beryl. The border zone contains quartz, potash feldspar, and "ruby" mica as essential minerals with accessory black tourmaline, garnet, and biotite.

In the western part of the pegmatite the central zone is separated from the upper border zone by a lens of quartz-perthite rock that is locally graphic in texture. At this place the central zone is thickest and is composed of a great assortment of complex minerals.

The maximum thickness and the mineral composition of the pegmatite units are shown in Table 1, and the structural relations of the different units are illustrated in Fig. 2. The occurrence of a quartz-perthite unit, locally graphic textured, separating a hanging-wall mica-rich zone and inner complex mineral zones, has also been reported for several granitic pegmatites in the United States (Cameron and others, 1949).

The hanging-wall border zone where best developed is about 1.5 meters thick. In its upper part, against the schist, mica books composed of "ruby" muscovite and intercrystallized biotite are arranged edgewise to the contact. In the lower part of the zone is a concentration of nests of randomly oriented "ruby" muscovite books and quartz, free of biotite, imbedded in potash feldspar. Quartz, tourmaline, and garnet are much more abundant in the upper part of the zone. The mica-rich part of the zone is called a "mica shoot."

The quartz-perthite unit has a maximum thickness of 4 meters. Perthitic feldspar makes up about 80 per cent of the rock and quartz and white muscovite the rest. Locally, particularly near the apex of the roll, or minor arches, the rock is graphic in texture. The largest individual cleavage surfaces in the feldspar are 1.5 meters across.

Irregular cavities are numerous in the central zone, particularly in the western, thicker part of the pegmatite. In addition to quartz and microcline, the mineral assemblage includes albite, tourmaline, beryl, cassiterite, tantalite, microlite, spodumene, cookeite, garnet, and muscovite. Altered spodumene "logs" up to 80 cm. long and albite "logs" up to 25 cm. long occur in the zone, and both minerals show effects of deuteric alteration and intense weathering.

The footwall border zone has the same essential mineral composition as the hanging-wall zone but differs in being thinner, in having less muscovite and black tourmaline, in lacking the conspicuous edgewise growth pattern of mica books at the contact, and in being relatively free of biotite.

The Mica Shoot

The mica shoot in the hanging-wall zone provides the principal source of commercial muscovite in the pegmatite. In the western part of the pegmatite, east of the "roll" (Fig. 3), the concentration of mica in the shoot approaches an average of 20 to 25 per cent by volume not all of which is commercial. The average thickness of the mica shoot in 1945 was 0.5 meter, and the calculated recovery of the commercial mica in the mica shoot was about 400 kilograms per cubic meter, or approximately 12 per cent by weight. Elsewhere in the hanging-wall zone the recovery of mine-crude mica ranges from 50 to 100 kilograms per cubic meter. The sheet mica trimmed from the Golconda mine-crude mica averages between 10 and 15 per cent by weight. Based on prices paid for sheet mica in 1943–45, the mine-crude mica at the Golconda mine was evaluated at approximately $180 per metric ton and the mica shoot itself between $20 and $30 per ton. This mica shoot thus has a higher value per ton of rock than the ore of most gold mines in Brazil, although its reserves, unfortunately, are much less. The economic features of the mica mines in Minas Gerais have been fully discussed (Pecora and others, 1950).

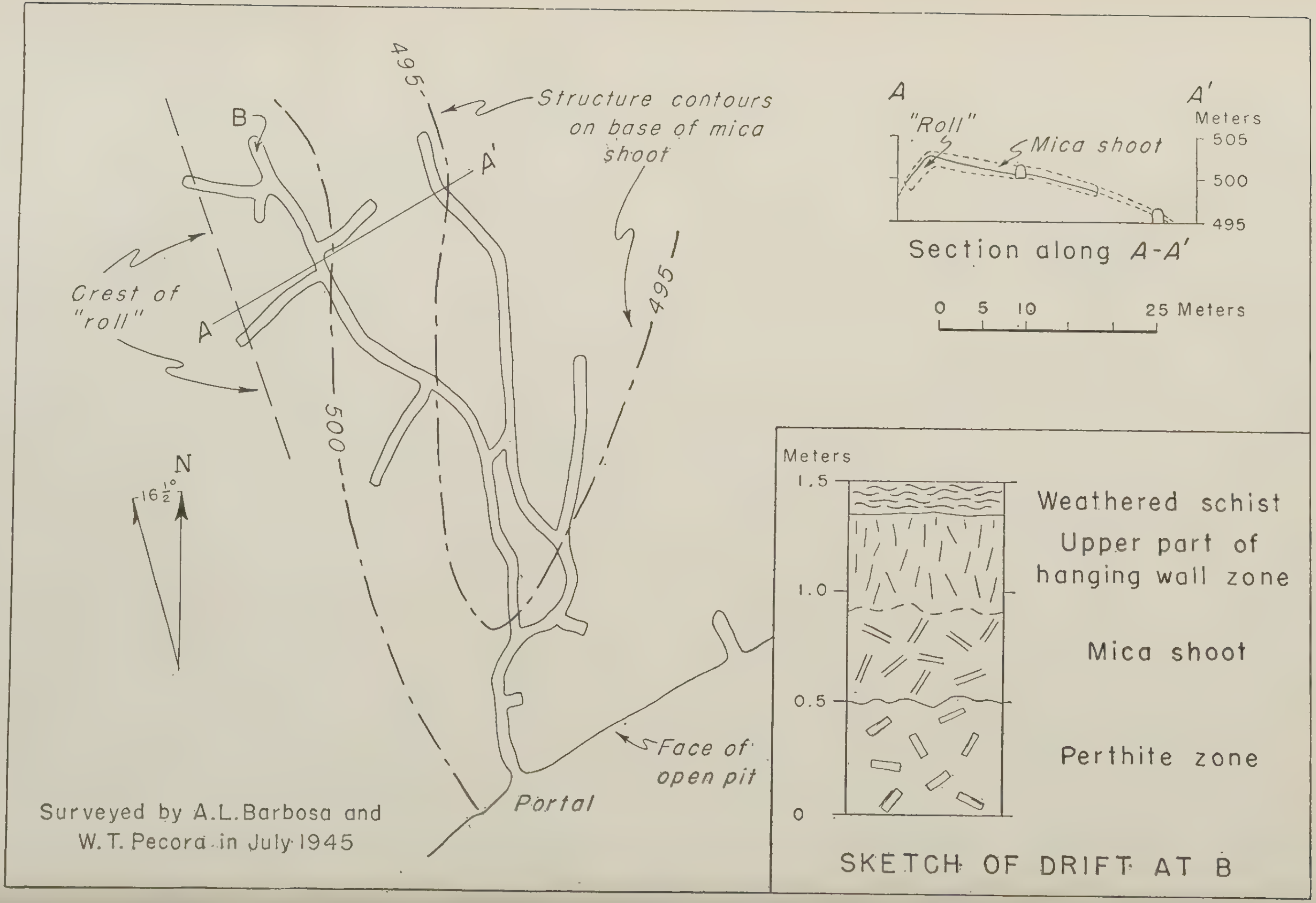
Structure contours on base of mica shoot
495
500
495
A
A'
B
Crest of "roll"
N
16½°
Face of open pit
Portal
Surveyed by A.L.Barbosa and W.T.Pecora in July 1945
A
A'
"Roll"
Mica shoot
Meters
505
500
495
Section along A-A'
0
5
10
25 Meters
Meters
1.5
1.0
0.5
0
Weathered schist
Upper part of hanging wall zone
Mica shoot
Perthite zone
SKETCH OF DRIFT AT B

TABLE 1. GENERAL DESCRIPTION OF THE INTERNAL UNITS OF THE GOLCONDA PEGMATITE, GOVERNADOR VALADARES, MINAS GERAIS, BRAZIL

Unit	Maximum thickness (meters)	Essential minerals	Accessory minerals	General comments
Hanging-wall border zone	1.5	Potash feldspar, quartz, and "ruby" muscovite	Biotite, garnet, and black tourmaline in upper part of zone	Muscovite books, with intercrystallized biotite, are arranged edgewise to contact in upper part of zone. Commercial sheet mica obtained only from lower part of zone
Quartz-perthite zone	4	Perthite and quartz	White muscovite	Quartz and microperthite are locally graphic, especially near the crest of rolls
Central zone	5	Potash feldspar, albite, and quartz	Beryl (morganite and aquamarine), tourmaline (green and blue), garnet (spessartite), cassiterite, tantalite, microlite, spodumene, cookeite, and muscovite	Cavities lined with accessory minerals are most common in the western part of the pegmatite
Footwall borzone	1	Potash feldspar, quartz, and "ruby" muscovite	Black tourmaline, garnet	Muscovite books too small sized to be commercial

DESCRIPTIVE MINERALOGY

Essential Minerals

Quartz and feldspar make up more than 95 per cent of the bulk of the pegmatite. Microcline perthite is the principal feldspar in the border zone and in the quartz-perthite zone. Because of the advanced state of weathering, the relative abundance of the different feldspars in the separate zones cannot be established with any assurance. In the central zone, nevertheless, albite is common and the microcline lacks any visible perthitic texture. The quartz-perthite rock is locally graphic and in the thickest part of the zone cleavage faces of the perthite measure up to 1.5 meters across.

Quartz is scattered throughout the border zone, but is much less abundant in the mica shoot and quartz-perthite zone than in either the border or central zones. Quartz lenses, ramiform quartz masses inter-grown with microcline, and cavities lined with clear or smoky quartz crystals are scattered throughout the central zone and are best developed in the thicker, western part of the pegmatite. The crystals range in size from a few centimeters to one-half meter, are commonly in parallel growths, and show steplike development of the prism and rhombohedron faces.

The term "ruby muscovite" is widely used in the mica industry for commercial muscovite that is yellow-brown or red-brown in color. In the Golconda pegmatite, ruby mica is distributed only in the border zone, whereas the white, green, and yellow varieties are scattered in the inner part of the pegmatite. Only the ruby muscovite occurs in large enough size and of quality to be of commercial value. Books of commermercial ruby muscovite are imperfect and rarely exceed 30 centimeters in breadth and 5 centimeters in thickness. Size and quality of the knife-trimmed sheet mica produced from the crude books is reported according to the India Standard, and a breakdown of about 25 tons of sheet mica recovered in 1944 from the mine is given below:

Size	*Area* (*sq. in.*)	*Kilograms*
6	1	12,282
5½	2½	3,311
5	3	5,533
4	6	2,504
3 and larger	10+	1,483
	Total	25,113

Quality		*Kilograms*
Good-Stained and better		12,986
Stained-A and worse		12,127
	Total	25,113

In the thick western part of the central zone, rounded masses of concentrically grown brown muscovite are embedded in the weathered feldspar. The curved planar surfaces are marked by bumps so that the Brazilian name of "mica de tatu" (armadillo mica) is not inappropriate.

Biotite is locally intergrown with the ruby muscovite in the upper border zone, but it is rare or absent in the commercially significant mica shoot.

Accessory Minerals

Garnet (spessartite)

Red-brown garnet is abundant as crystals less than 1 centimeter in size in the outer part of the border zone and in the schist of the country rock. In the central zone, however, irregular masses of dark brown garnet up to 15 centimeters across are embedded in weathered feldspar. Owing to the extent of alteration, the garnet masses now consist of clear brown fragments up to 5 millimeters in diameter in a matrix of black fine-grained oxide of manganese. The dark brown garnet was determined to be spessartite on the basis of the following properties: index of refraction = 1.808 ± 0.005; specific gravity = 4.18 ± 0.05; unit cell size $(a) = 11.57 \pm 0.01$A; $MnO = 14.16\%$.

These data were furnished to S. B. Levin of the Squier Signal Corps Laboratory, Fort Monmouth, N. J., and using the method he devised (Levin, 1949) he calculated the following components of the Golconda spessartite:

Components	*Per cent*
Almandite	54
Spessartite	32
Pyrope	8
Grossularite	6
Andradite	0

Cookeite

Pure masses of cookeite are intergrown with other rare minerals or have formed on surfaces of crystals in the cavities. Some logs of spodumene up to 40 centimeters long are completely replaced by very fine grained cookeite. Individual aggregates in other associations are up to 2 millimeters in size. The colors of the mineral include rose, lavender, and pale lilac, and the material is composed of hexagonal plates and sections.

A mass of pale lavender dense material was purified for special study. The optical properties of the mineral are:

$$\left.\begin{array}{l}\alpha=1.578\\ \beta=1.581\\ \gamma=1.598\end{array}\right\}\pm 0.002 \qquad \begin{array}{l}\text{Biaxial }(+)\\ 2V=45°\text{, calculated}\end{array}$$

A partial spectrochemical analysis of this material by A. T. Myers gave the following results: $Li_2O = 2.7\%$; $K_2O = 0.0X$; $Na_2O = 0.0X$; and $Cs_2O = 0.0X$. Semiquantitative chemical analyses indicate that some Golconda cookeite may have up to 4 per cent Li_2O.

In habit, optical properties, and *x*-ray powder pattern the Golconda material resembles cookeite examined by the authors from other mineral localities, and its chemical composition is like that of cookeite reported by Brammall and others (1937). Quensel (1937) has described cookeite that has formed by replacement of spodumene in a pegmatite from Sweden.

Beryl (Morganite and Aquamarine)

Pale pink beryl (morganite) was found associated with quartz, albite, and cookeite in the central zone. A number of specimens up to 20 centimeters in diameter were recovered. The beryl occurs in irregular crystalline aggregates of many deeply etched parallel and sub-parallel individual crystals. Many crystal faces are represented, but no complete individual crystals were found and the crystals are highly malformed. Individual crystals are up to about 5 centimeters in size. Much of the beryl is free from flaws and inclusions, but it is too pale in color to be classed as good gem material.

The pink beryl has the following properties:

Uniaxial (−) $\left.\begin{array}{l}\omega=1.591\\ \epsilon=1.584\end{array}\right\}\pm 0.002$

Sp. gr. $=2.78\pm 0.01$

These values indicate a beryl very high in alkalies, according to unpublished data of W. T. Schaller.

The pink beryl crystals have an unusual crystal habit in that they are tabular parallel to $\{10\bar{1}1\}$.

Bluish-green beryl (aquamarine), in part of gem quality, was also found as a rare constituent of the central zone.

Cassiterite

Coarsely crystalline masses of cassiterite were encountered in the central zone, associated with other rare minerals. The well developed crystals are as much as 4 centimeters in diameter. They are usually subhedral but a few rough euhedral crystals were found showing only faces of the unit dipyramid.

The color of the cassiterite in some zoned crystals ranges from black to greenish yellow. Spectrochemcial analyses by A. T. Myers of each of these zones in a single crystal are given in the table below.

The presence of columbium and tantalum in Golconda cassiterite is

	Black Cassiterite	Greenish yellow Cassiterite
MnO	0.65%	0.14%
Fe_2O_3 (total)	0.29	0.06
SnO_2	X0.0	X0.0
Cb_2O_5	0.46	0.18
Ta_2O_5	1.0	0.1
Si	0.X	0.X
Al	0.0X	0.00X
Mg	0.0X	0.00X
Ca	0.00X	0.000X
Cu	0.000X	0.0000X
Ti	0.000X	0.00X

Not found: Ag, Au, Be, Bi, Co, Cr, Ga, Ge, In, La, Li, Mo, Na, Pb, Sb, Th, Tl, V, Y, Zn, Zr.

noteworthy. Edwards (1940) likewise reports solid solution of columbium and tantalum in cassiterite from Australia, and other localities showing similar relationship are listed in Dana's *System of Mineralogy* (7th ed., p. 576). Borovick and Gotman (1939) demonstrate from a compilation of spectrochemical data of numerous cassiterite samples that cassiterite from pegmatites generally contains more columbium than cassiterite from veins.

Manganotantalite

Manganotantalite occurs as small, well formed, prismatic crystals associated with microlite in cavities of cassiterite. The crystals are as much as 1 centimeter in length by 2 millimeters in diameter. In color they range from black to reddish brown.

Unit cell dimensions[4] determined on one brown crystals by the Weissenberg method are as follows:

$$\left.\begin{array}{l} a = \ \ 5.14\ \text{A} \\ b = 14.44\ \text{A} \\ c = \ \ 5.78\ \text{A} \end{array}\right\} \pm 0.01$$

The specific gravity as determined by the average of several crystals measured on the Berman balance is 7.34 ± 0.05.

The Golconda manganotantalite crystals are elongated parallel to [001] and are prismatic to acicular in habit (Fig. 4). The prism zone is usually striated, and the terminations are simple, usually consisting only of a basal pinacoid. Termination of some crystals by a single face of the

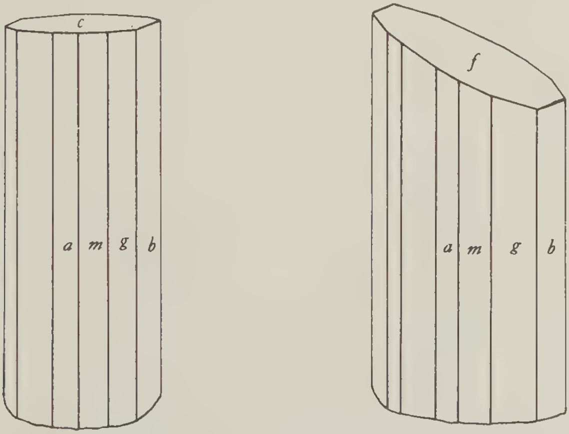

FIG. 4. Crystal drawings of manganotantalite from the Golconda pegmatite.

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[4] Unit cell dimensions are given here in the same orientation used in Dana's *System of Mineralogy*, 7th ed., rather than in the conventional form $c<a<b$. The same orientation is followed in the following section on morphology.

dome *f*{032} causes them at first glance to appear monoclinic. The forms observed by measurement of several crystals were: {001}, {010}, {160}, {130}, {110}, {032}, and {131}.

A spectrochemical analysis of the Golconda manganotantalite by A. T. Myers gave the following results:

MnO	14.0 %
FeO (total)	0.52
SnO_2	0.96
Cb_2O_5	14.0
Ta_2O_5	71.0

Ti 0.X; Si, Mg 0.0X; Ca 0.00X; Al, Cu 0.000X.
Not found: Ag, As, Au, Be, Ge, La, Li, Th, Tl, U, Zn, Y.

In Dana's *System of Mineralogy* (7th ed., pp. 782–785) the usage of manganoan and ferroan as adjectival modifiers for tantalite and columbite is not completely in accord with the original suggestions of Schaller (1930). The term manganoan tantalite should be used for a tantalite in which the iron content is in excess of the manganese and not, as used in the Dana volume, for a manganese-rich tantalite. The iron end member is implied in the single name of tantalite or columbite, whereas the manganese end member is implied in the name manganotantalite or manganocolumbite.

Microlite

Microlite was found in small amounts as honey-yellow crystals associated with cassiterite and manganotantalite in the complex mineral zone. The crystals show rough octahedral form, and their maximum size is about 3 millimeters.

Some of the physical properties of the microlite are as follows: index of refraction = 2.055 ± 0.005 (selenium melts); specific gravity = 6.12 ± 0.05; unit cell (*a*) = 10.422 A ± 0.001.

The Golconda microlite is very near the pure tantalum end member of the microlite-pyrochlore series as indicated by its index of refraction, specific gravity, and the spectrochemical analysis (by A. T. Myers) given below:

MnO	0.42%
FeO (total)	0.22
SnO	1.9
Cb_2O_5	1.9
Ta_2O_5	X0.0
CaO	X.0

Na, Si, Ti 0.X; Mg 0.0X; Al, Bi, Pb 0.00X; Ag trace.
Not found: As, Au, Be, Co, Cr, Ge, Ga, La, Li, Mo, Th, Tl, U, V, Y, and Zn.

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DETERMINATION OF ALL PRINCIPAL INDICES OF REFRACTION ON DIFFICULTLY ORIENTED MINERALS BY DIRECT MEASUREMENT

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Abstract

A device and method of general applicability is described whereby one can determine all principal indices of refraction on a single grain of non-opaque substance by direct measurement. The materials used are easily obtained and cost less than one dollar.

Theory

In the course of examining the optical properties of amphiboles, carbonates, and micas during a recent petrographic investigation, it occurred to the writer that no rapid and, at the same time, rigorous systematic method existed for the accurate determination of all the principal indices of refraction on minerals which are difficult to orient due to perfect cleavage at undesirable angles to the optical directions. In attempting to measure γ and α on a hornblende, the writer experimented with a grain mounted on the end of a needle capable of rotation in the plane of the microscope stage. It soon became apparent that any linear direction within any mineral grain could be turned into the plane of the microscope stage, and that by rotation of the microscope stage this linear direction could be rotated into any azimuthal direction (i.e. parallel to the north-south cross hair). The principal indices of refraction represent linear directions within any non-opaque mineral grain, and therefore can be turned into parallelism with the stage of the microscope parallel to the north-south cross hair, the measuring position. There remains to be be recognized a proper orientation of one of the principal indices when it is attained. It can be stated as a result of the theory of the optical indicatrix that whenever one of the principal indices of refraction is in the measuring position, an interference figure will show, in the general case, an isogyre along the east-west cross hair. This can readily be visualized by studying a skiodrome model. Conversely, whenever a dark bar occurs along the east-west cross hair under conoscopic observation, one of the principal indices of refraction lies in the measuring position parallel to the north-south cross hair.

The Device

The following materials are desirable for the construction and use of the device:

1 object glass (1 inch × 3 inches × .04 inch).
1 scrap object glass (preferably .06 inch thick).

1 scrap object glass (preferably .04 inch thick).
waterglass (in small dropper bottle with rod applicator).
1 diamond point or other suitable device for cutting glass (corundum satisfactory).
1 2.5 inch corsage pin with spherical knob on end (a standard size available at almost any floral shop). Pin must be straight.
Beeswax
1 12 mm. circular or square glass cover slip.

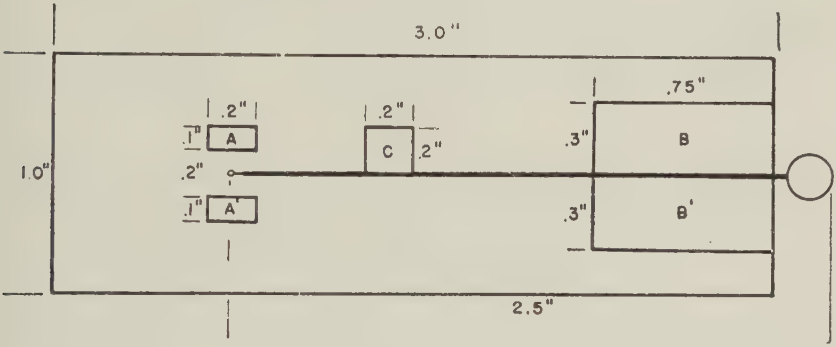

FIG. 1

In Fig. 1 a top view of the device is shown. *A* and *A′* represent pieces of the .04 inch object glass cemented with waterglass to the slide. These serve as support for the cover slip. *B* and *B′* represent guides for the needle and should be cemented to the slide with the needle in position in order to get a snug fit. They should be made out of .06 inch object glass. *C* represents a piece of .06 inch object glass beveled so as to overlap the needle with light pressure. It should be installed with the needle in position after *A*, *A′*, *B*, and *B′* have been mounted. Its purpose is to prevent the needle from flying up when the knob is adjusted.

MOUNTING THE GRAIN

One of the more difficult problems in developing the device involved a practicable method of mounting the mineral grain on the end of the needle with a cement which would not be attacked by the immersion media. It was found that a grain could be picked up on the end of the needle by first making the point sticky with a minute amount of beeswax. Inasmuch as the beeswax is attacked by most immersion media, it is necessary to further armour the point of contact between the grain and the needle, being careful not to coat the grain. This is done with waterglass by carefully twirling the point of contact of the needle and grain on the topside of the waterglass rod applicator. A very small amount of waterglass will suffice for a firm connection. Waterglass has the desirable property of drying very rapidly. Mounting is facilitated by the use of a hand lens or binocular microscope. After a little practice it becomes very simple to mount even relatively small grains in this fashion. Another method which is occasionally easier for mounting small grains involves cementing with waterglass a quarter inch segment of fine capillary tubing on the end of the needle. After the cement has hardened, the end is moistened with a small amount of waterglass and immediately touched to the mineral grain, which adheres rather firmly. The capillary tubing helps to eliminate the tendency of the waterglass to "ball up" near the end of the needle. Both methods have been used with success by the writer.

PROCEDURE

After the grain is mounted, the needle is placed on the slide as shown in Fig. 1. The cover slip is placed on the cover slip supports, and an appropriate immersion medium is allowed to be drawn into the immersion chamber by capillarity. Using a 4 mm. objective (preferably of N. A. .65 with longer working distance than the N. A. .85 objective) the grain is viewed under conoscopic conditions. By rotation of either the needle or the stage or both, an isogyre is brought into the field of view and oriented along the east-west cross hair. One of the principal indices of refraction is oriented parallel to the north-south cross hair in the measuring position. Maintaining this orientation and using the spherical knob on the end of the needle as a spherical projection, two ink dots are placed on the knob so that the line between them is parallel to the north-south cross hair. The stage is then rotated 90°. By synchronous rotation of stage and needle, the line between the two ink dots is rotated in a plane perpendicular to the north-south cross hair until a second isogyre lies along the east-west cross hair. A second principal index of refraction lies in the measuring position parallel to the north-south cross hair. Two more ink dots are placed on the knob parallel to the north-south cross hair. The third principal index of refraction lies perpendicular to the plane containing the other two, and can be immediately oriented, by synchronous rotation of stage and needle, into parallelism with the north-south cross hair, in which position a third isogyre will be seen to lie along the east-west cross hair. Ink dots are placed on the knob as before. The purposes of marking the knob are merely to prevent duplication of steps, to give an approximately correct orientation of the indicatrix in the attached mineral grain, and to aid in achieving a rigorous sequence of steps. Accurate orientation in each case is based on the presence of an isogyre along the east-west cross hair.

With the information concerning principal vibration directions recorded on the knob of the needle, it is a relatively simple matter to determine the indices parallel to these directions by immersing the grain in successive media and orienting in accordance with the above method. A medium may be removed by adsorption into some absorbent paper, followed by one flush of the next medium to be used. Dilution effects thus become practically negligible. It also becomes unnecessary to move from a desired orientation. If it is desired to do accurate work, the cell may be flushed with acetone. It is then necessary to wait for the slide to warm to room temperature because of the cooling caused by the evaporation of acetone.

The uniaxial mineral represents a special case of the biaxial mineral where either α and β are equal, or γ and β are equal. The same procedure therefore applies to uniaxial minerals. The uniaxial character will be demonstrated by the fact that either α and β will be found equal (positive mineral) or γ and β will be found equal (negative mineral).

CONCLUSIONS

This method has been used by the writer and his students with considerable success. It offers little or no advantage in speed for grains of minerals which tend to assume random orientation in powdered form. However, it has been found to be practicable in the cases of minerals which tend to assume preferred orientations at undesirable angles. For this latter class of minerals the method is believed to have the following advantages:

1. Rigor. No assumptions need be made about cleavage angles, as the method relies entirely on functions of the optical indicatrix.
2. Speed. The writer has found that, by using this method, he is able to determine all the principal indices on a clinoamphibole just about as rapidly as he can on an olivine, using conventional methods.
3. Teaching aid. It has proved a valuable aid in teaching elementary students to visualize the orientation of the indicatrix within crystals.
4. Simplicity. It is unnecessary to carry out any trigonometric calculations.
5. Use of only one grain of mineral. This has an added advantage if impurities are abundant in the source material, as in many sediments.

ACKNOWLEDGMENTS

The writer is deeply indebted to the teaching, and the wise, sympathetic counsel of Professor Larsen. The writer further wishes to acknowledge the help received in testing the proposed method from the students in his elementary optical mineralogy class at Missouri School of Mines and Metallurgy. Lawrence Blade critically read the manuscript. Drafting was done by John Emery.

THE DARK-FIELD STEREOSCOPIC MICROSCOPE FOR MINERALOGIC STUDIES*

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ABSTRACT

Dark-field illumination has not been widely used in the study of minerals, but the stereoscopic microscope can readily be adapted to make use of the dark-field effect. The dark-field stops that may be used with the stereoscopic microscope are described, together with some of the applications for which the method has proved most useful.

Dark-field illumination as applied to the stereoscopic microscope, commonly known as the binocular microscope (Greenough type), permits unusually intense illumination, gives strong stereoscopic effects, and reveals internal structures in mineral grains, all of which may be helpful in mineral studies. The method of equipping a microscope with dark-field illumination and some of the types of study that are facilitated by such illumination are described below.

The stereoscopic microscope, (commonly called the Greenough type), equipped with a condenser system for giving dark-field illumination, has many applications in mineral investigations. A description of the equipment and some of its applications may be of general interest. The only modifications of the stereoscopic microscope are the addition of a source of transmitted light and a dark-field condenser. The writer has been using the simple Carl Zeiss plankton condenser. The type of condenser supplied with an inexpensive biology microscope and dark-field stop may be used. This dark-field stop gives the same effect, but the plankton condenser has the advantage of being effective through a greater working distance. A source of transmitted light is required and so the lower stand of an old biology microscope serves very well if the tube and its support are removed and the dark-field stop used or the plankton condenser substituted. If a stereoscopic microscope with stage and mirror is available it may be adapted to hold the condenser system.

A permanent set-up is desirable. Therefore, it is advisable to have the stand carrying the condenser and mirror permanently fixed in the proper position under the microscope. Any good stereoscopic microscope is suitable, but ample clearance is necessary. Paired objectives giving somewhat higher magnification than commonly used are desirable with dark-field illumination and the combinations giving magnifications of 24, 48, and 72 times are found most suitable.

The blackness of the field permits very intense illumination of the objects that reflect light into the field of view. For most work, a 500-watt incandescent projection bulb does not produce undue eyestrain. The intense illumination tends to make even very small objects or structures clearly visible.

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With dark-field illumination, light is reflected from all surfaces of discontinuity, thus bringing out all boundaries of grains, inclusions in grains, or internal structures marked by slight differences in indices of refraction. Color differences also show very clearly. Any dust or other foreign material within the field is brought out with equal intensity. For this reason, careful cleaning of grains and the glass slide is imperative. If grains are mounted in immersion media, the upper surface of the cover glass must be free from dust and smears.

Dark-field illumination has been effective in making counts involving the ready recognition of a number of minerals. For the support of the grains, a glass slide 8 by 10 centimeters in size is prepared with a number of very fine parallel scratches made with a diamond. Only those grains in contact with a scratch are counted. A magnification of 48 times is most commonly used. In general, counts can be made much more rapidly under the stereoscopic microscope than with the petrographic microscope, but the latter may be needed for preliminary identification. Color and form show so clearly that these are commonly adequate for ready recognition.

A study, involving the use of dark-field illumination with the stereoscopic microscope, was made on sands which formed the ballast material for the incendiary balloons that the Japanese dispatched to this continent in great numbers during World War II. The objective of the study was to locate the launching sites of the balloons, but the sands themselves had unusual mineralogic interest. The method used in this study will serve to illustrate the general method. The mineral composition of the sands was very constant, but the proportions varied within wide limits. A representative composition determined by count and corrected for specific gravity is shown on the following page.

The grains were evidently from clean beach sands and did not require washing. The mineral grains may have been somewhat rounded. However, the euhedral form of some of the crystals indicated that rounding by erosion was slight for most of the material. The magnetite, which was abundant in some of the samples, was separated by a hand magnet and weighed; the remaining grains were essentially transparent minerals. The dark-field stereoscopic microscope was especially effective in rapidly distinguishing hypersthene from augite. Under the petrographic microscope either the pleochroism or the extinction of the hypersthene had to be recognized by rotation of the microscope stage. Under unpolarized light of the stereoscopic microscope, hypersthene was some shade of brown

TABLE OF MINERAL COMPOSITION OF BALLAST SANDS OF JAPANESE BALLOONS

	Per Cent
Hypersthene	52
Augite	7
Hornblende	8
Garnet	1
Magnetite	10
Quartz phenocrysts	8
Quartz, detrital	8
Plagioclase (An_{30}-An_{50})	6
Plagioclase (An_{90}-An_{100})	Sparse
Zircon	<1
Sphene	Sparse
Biotite	Rare
Hornblende schist	Present
Shell material	Not included
Volcanic glass	Rare

and the augite was slightly greenish. The hornblende ranged from dull brown to dark green and commonly its cleavage was easily observable. The quartz phenocrysts ranged from sharply euhedral to partly rounded and were readily recognizable as having the form of high-temperature quartz. This form also served to distinghish them from the detrital quartz. The intermediate plagioclase was glassy clear; the plagioclase, hypersthene, augite, and euhedral quartz were evidently of volcanic origin, as were the sparse glass shards. The hornblende, the biotite, and probably the garnet were derived from metamorphic rocks, as the hornblende in individual grains and that in the greenstone fragments were similar. The source of the magnetite was not evident from the microscopic study. The detrital quartz seems to have been derived from sedimentary rocks.

Photomicrographs shown in Plate 1 fail to do justice to the method, as the stereoscopic effect is lacking. Figure (*a*) shows the euhedral form of the hypersthene. Figure (*b*) shows the quartz phenocrysts with some of their inclusions. In this photograph the field is light instead of dark, owing to a milky-glass background, which gave diffused light and made the quartz stand out in greater relief. Even without the binocular effect the euhedral form is shown.

A liquid inclusion with vapor vacuole in a much larger quartz crystal from Arkansas, having the morphology of low-temperature quartz is shown in Fig. (*d*). The figure shows a negative quartz crystal. The low-temperature form and the gas bubble floating in the liquid are visible even without the binocular effect.

Figure (*c*) represents a photograph of a specimen of thread pumice (reticulite) collected by Chester K. Wentworth at Kilauea Volcano,

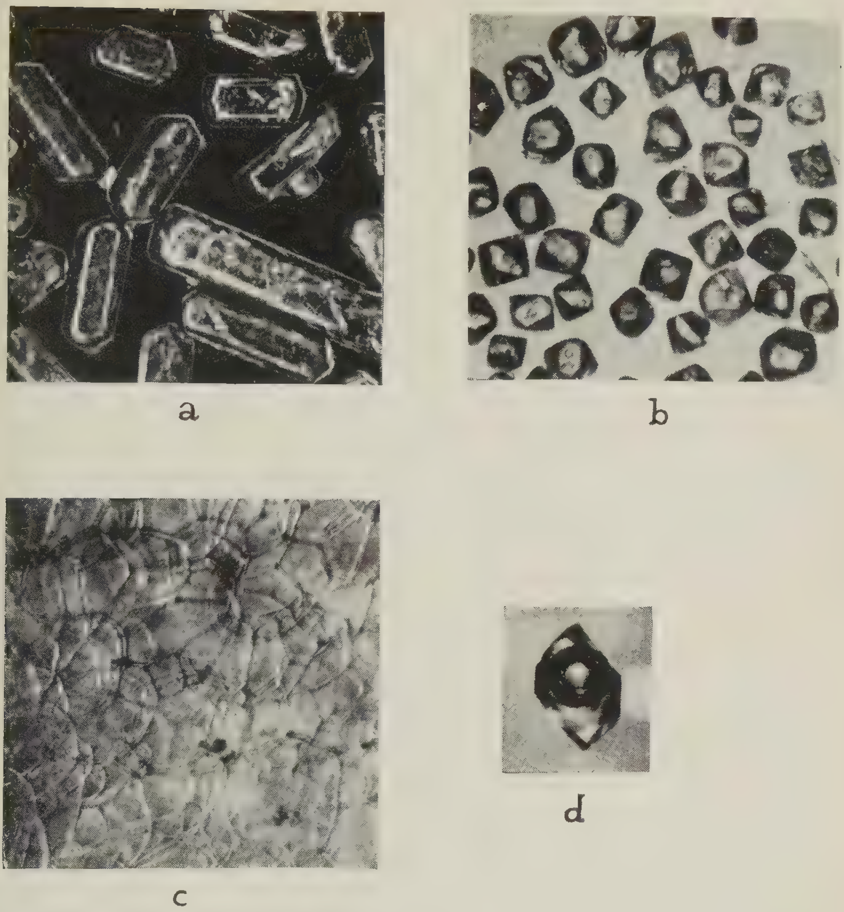

PLATE 1

Photo Micrographs Using Dark-field Illumination.

Island of Hawaii. The specimen was impregnated with plastic and the surfaces polished. The reticulate threads of basaltic glass are well shown.

At times it is helpful to supplement the study of thin sections under the petrographic microscope by transferring them to the dark-field microscope. Minute traces of oxidation products of iron-bearing minerals and of films of leucoxene show up clearly, and even sparse gibbsite disseminated in kaolin produces conspicuous areas of greater opacity. The intense illumination allows light to be transmitted by materials that appear opaque under normal illumination; and so significant colors may be observable.

The stereoscopic microscope with dark-field illumination gives the same brilliant index of refraction effects which have been described by Nelson B. Dodge.[1] This effect as seen with the equipment here described has not been used as a general means of mineral identification. However, transparent minerals with slightly different indices of refraction may be distinguished at a glance. Quartz mounted in an immersion medium with intermediate index, is bordered by blue or by red and blue, and sodic plagioclase is bordered by red and orange. However, the differences in indices of refraction are hardly as clear-cut as they would be with good inclined illumination under a petrograhic microscope.

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[1] Dodge, Nelson B., The dark-field color illumination method: *Am Mineral.*, **55**, 541–549 (1948).

MISERITE FROM ARKANSAS; A RENAMING OF NATROXONOTLITE*

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Abstract

A reinvestigation of a pink silicate mineral, occurring with and formed from wollastonite in a contact metamorphosed shale at Potash Sulphur Springs, Ark., has shown that the original analysis is in error. Because of the high sodium content (4½% Na_2O) reported and the supposed relation of the pink mineral to xonotlite, the name natroxonotlite was suggested by J. Francis Williams in 1891 in the famous volume 2 of the Arkansas Geological Survey. New analyses show that the dominant alkali is potassium, and not sodium, and dehydration and *x*-ray studies indicate that the mineral is not a xonotlite. The name natroxonotlite, therefore, is "glaringly false in signification" and the mineral is renamed miserite after Dr. Hugh D. Miser, geologist of the U. S. Geological Survey.

The original analysis indicated that miserite was intermediate in composition between wollastonite and pectolite, but the present investigation shows that this is not so. The formula of miserite is $KCa_4Si_5O_{13}(OH)_3$ or $K_2O \cdot 8CaO \cdot 10SiO_2 \cdot 3H_2O$. Indices of refraction 1.58 to 1.59. $G = 2.84$. Biaxial, positive, 2V small. The dehydration curves of miserite and of xonotlite are not similar. *X*-ray powder patterns of miserite are different from those of wollastonite, pectolite, and xonotlite.

Introduction

Some years ago, while compiling the available published information on wollastonites containing small but appreciable amounts of manganese or of iron, in order as far as possible to complete the data on the wollastonite-bustamite series, I noted the analysis of a white wollastonite from Arkansas, with 2.03 per cent of FeO and 2.08 per cent of MnO, published by Williams (1891, p. 356) and listed in the Supplement to Dana's System of Mineralogy, 6th ed., page 1052.[1] Associated with this white wollastonite was a pink calcium silicate (miserite), reportedly with less iron and manganese but with more water and nearly 4½ per cent of soda. Williams called it a pink variety of wollastonite, related to xonotlite, the best-known hydrated calcium silicate at that time; and because of the sodium supposed to be present, he suggested the name natroxonotlite for it. An analysis, by R. N. Brackett, placed the mineral halfway between wollastonite and pectolite in composition.

The published description of the pink mineral suggested a number of intriguing relationships between these minerals, and as no specimen of it could be located, I welcomed an opportunity in November 1937 to visit the Arkansas locality in company with Hugh D. Miser and Clarence S. Ross. Abundant material was collected, and on return to Washington, some of its properties were determined. These were found not to be inter-

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[1] I am still looking for wollastonites containing 1 to 3 per cent MnO.—W.T.S.

mediate between the similar properties of wollastonite and pectolite. A reexamination of the pink mineral showed that the old analysis was in error. The pink mineral is not intermediate between wollastonite and pectolite, it is not a sodium (natro) compound, and it is not a xonotlite. The name natroxonotlite is, therefore, "glaringly false in signification"[2] and so grossly misleading as to be untenable and a new name is called for. No name could be more appropriate than miserite,[3] after Dr. Hugh D. Miser, born in Pea Ridge, Ark., geologist of the U. S. Geological Survey since 1907, who has contributed so much to our knowledge of the geology and mineralogy of his native state. According to Dr. Miser, practically all of the available specimens of miserite have now been collected and deposited in the U. S. National Museum in Washington.

New analyses show that potassium and not sodium is the essential alkali, and the formula derived for miserite is $KCa_4Si_5O_{13}(OH)_3$ or $K_2O \cdot 8CaO \cdot 10SiO_2 \cdot 3H_2O$. Through the courtesy of Prof. James D. Burfoot, Jr., of the Department of Geology, Cornell University, I was able to examine Williams' type specimen of his pink wollastonite or natroxonotlite, with which our material is identical. Blowpipe tests on many fragments of the type material gave the characteristic violet flame coloration of potassium. Only an occasional small fragment of the pink mineral gave a strong yellow sodium flame, but every such piece was seen to contain inclusions of aegirite. Similar results were obtained on our material.

OCCURRENCE

Miserite occurs at Potash Sulphur Springs (now called Wilson Mineral Springs) about halfway between Hot Springs and Magnet Cove, Ark. It is a constituent of the (quoting Williams) "very beautifully banded metamorphosed shales"[4] in which "occur the beautiful pink [miserite] and white wollastonite crystals which were exposed when the rock was blasted open in building the road" [from Lawrence Station to the Springs hotel. The boulders in the creek bed from which we collected are undoubtedly pieces resulting from this blasting. The masses of white wollastonite probably represent calcareous portions of the original shale—W.T.S.]. The metamorphosed and indurated shale, now composed essentially of orthoclase, aegirite, wollastonite, and miserite, is in the contact band crossing the road about 700 yards south of the hotel, and was

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[2] Dana, E. S., System of Mineralogy, 6th ed., 1892, Introduction, IV. Nomenclature. 13. Limitations of the Law of Priority, b. "When it [the name] is glaringly false in signification," p. xliii.

[3] Pronounced my′-ser-ite.

[4] This rock polishes well; and the contrasting pink miserite, white wollastonite, and greenish aegirite present a very attractive appearance. It is to be regretted that more material is not available.

"formed by the intrusion of a vertical dike of eleolite syenite 30 feet wide, cutting the shales in a nearly east and west direction. The shale on both sides of this dike is of a greenish gray color and but little decomposed while the igneous dike has almost diasppeared."

"Near the line of contact the shale has entirely lost its shaly, laminated structure and has the appearance of a glassy igneous rock. It is cut by numerous small sheets and masses of pink [miserite] and white [wollastonite] minerals which present a strong contrast to the greenish color of the rock. . . . This white mineral [wollastonite] is generally covered with a layer of the pink scales [miserite]. . . . The pink variety of wollastonite [miserite] occurs in thin plates and scales. The color is apple-blossom-pink to pale rose color and is characteristic of the thin scales of this mineral.[5] It [miserite] might be considered as a xonotlite in which one-tenth of the lime (CaO) is replaced by soda (Na_2O) and in which the ratio of the normal silicate to the water is as 5:1 in place of 4:1 as in xonotlite. In view of this fact, the name *natroxonotlite* is suggested for this variety. . . . The natroxonotlite is formed by a decomposition of wollastonite. . . . "

Williams' description of miserite would indicate that it is in plates or scales, whereas it is fine-grained fibrous, and only the compact aggregates form slightly curved, scaly masses several inches in diameter but rarely more than a quarter of an inch thick. It commonly is in the white wollastonite, at many places forming a border zone between the wollastonite and the greenish host rock, which is much seamed with veinlets of wollastonite with or without miserite, and of aegirite.

An examination under the microscope of the wollastonite fragments (100 mesh), separated from the miserite by heavy solution, leaves little doubt as to the derivation of miserite from wollastonite, as stated by Williams. Many of these fragments of wollastonite are penetrated to a varying degree by long, extremely slender prisms of miserite, which are heterogeneously oriented or roughly parallel to the elongation of the wollastonite fragments and hence, when abundant, form aggregate pseudomorphs after wollastonite. Very rarely are the single long prisms of miserite wide enough to permit satisfactory optical observations. Disseminated small particles of orthoclase may be considered as evidence of the source of the potassium in the miserite. The ratio of Na_2O to K_2O in miserite is similar to that in orthoclase in general. A few fragments of miserite, carried down by the wollastonite, in the purification of the sample by use of bromoform, contain inclusions of smaller fragments of wollastonite, considered to be residual.

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[5] On surfaces of the rock which have been exposed to the weather for half a century, the miserite has taken on a more lavender hue.

OPTICAL PROPERTIES

The fineness of the fibers of miserite make it difficult to determine the optical properties satisfactorily. Most of the short prismatic pieces, as observed in crushed fragments, are compact aggregates of fine fibers, only approximately parallel. Occasional single unit fibers have parallel extinction with negative elongation; some of them are slightly more pinkish normal to the elongation. The axial plane is parallel to the elongation with an apparently centered acute bisectrix, with 2E measured about 60°. The sign is positive and the weak dispersion is $r>v$. Parallel to the elongation the index of refraction is close to 1.580 and normal thereto close to 1.590. Apparently orthorhombic.

Ignited miserite has indices of refraction of about 1.60–1.61±. Undetermined crystals (pseudowollastonite?), heterogeneously oriented in the fibers of ignited miserite, have developed. Ignition seems to produce a two-phase product.

The separated silica, resulting from HCl decomposition, without gelatinization, retains the size and shape of the original grains. This separated silica is weakly birefringent, has parallel extinction with positive elongation, and indices of refraction of about 1.41±.

CHEMICAL COMPOSITION

Selected pieces of the metamorphosed shale, richest in miserite, were ground to 100 mesh, washed dust free, and run through the Frantz isodynamic separator to remove the aegirite. Most of the orthoclase and wollastonite were removed by the use of proper heavy solutions (bromoform). The analyzed sample retained about 5 per cent of orthoclase and 3 to 5 per cent of wollastonite, as indicated by grain counts under the microscope.

Two analyses of the same sample of miserite were made, one by fusion with sodium carbonate and one by hydrochloric acid decomposition. The results of the analyses are shown in Table 1.

The recalculated values for the analysis by Na_2CO_3 fusion are obtained after deducting 5 per cent of orthoclase (the insoluble from the acid decomposition sample) and 1 per cent of sphene (based on the difference in percentages of TiO_2 in the two analyses) and recalculating to 100 per cent.

These results show several important differences from those obtained by Brackett (in Williams, 1891, p. 357). The most important one is that the essential alkali is potassium and not sodium, and hence the mineral name cannot be prefixed with "natro." Nearly twice as much water (all high temperature) as originally given is present. The percentages of FeO and of MnO earlier reported are much too high. If his deter-

TABLE 1. ANALYSES OF MISERITE FROM ARKANSAS

	By fusion with Na_2CO_3		By acid decomposition	
	Analysis	Recalculated	Analysis	Recalculated
Insoluble	—	—	5.27[3]	—
SiO_2	49.83	49.27	46.84	49.24
TiO_2	0.84	0.46	0.38[4]	0.40
Al_2O_3	1.62	0.74	0.84	0.88
Fe_2O_3	0.42	0.45	0.36[5]	0.38
FeO	0.06	0.06	0.06[6]	0.06
MnO	0.26	0.28	0.26[7]	0.27
CaO	34.76	36.69	34.61[8]	36.38
MgO	None	None	None	None
Na_2O	0.91[1]	0.97	0.89	0.94
K_2O	6.37[1]	5.88	6.00	6.31
H_2O-	None	None	(None)	(None)
H_2O+	4.89[2]	5.20	4.89[6]	(5.14)
	99.96	100.00	100.40	100.00

G. 2.84.

[1] Average of three determinations by different chemists. See text.

[2] Ignition loss. Average of 4.91 and 4.87. Total loss found by Nutting in his dehydration determinations, stopping at 980° C. = 4.53. Penfield tube method was unsuccessful, only 1.23 per cent being obtained, but an additional 3.27 per cent loss was found by blasting the removed sample in a platinum crucible, giving a total of 4.50. On separate sample of a scaly mass of fibers, an ignition loss of 5.26 was found.

[3] Probably slightly high (a few tenths of a per cent) as insoluble was dried only to 110° C for subsequent microscopic examination.

[4] Average of 0.38, 0.37.

[5] Average of 0.33, 0.39.

[6] Values taken from those determined on the fusion sample.

[7] Colorimetric. Average of 0.24, 0.26, 0.27.

[8] Average of 34.63, 34.58.

minations of iron and manganese were made gravimetrically, it is possible that the undetermined alumina and titania may be the reason for his high results. Brackett did not report either TiO_2, Al_2O_3, or Fe_2O_3. The associated white wollastonite was found to contain 0.50 per cent of FeO and 0.68 per cent of MnO, values considerably higher than found for the pink miserite, suggesting that the pink color may be due to titanium rather than to manganese. Comparison of hand specimens shows the colors of miserite, particularly on weathered surfaces, and of dumortierite from Dehesa, Calif., to be similar. I had previously suggested (Schaller, 1905, p. 101) that the lavender color of the dumortierite from Dehesa was due to titanium.

The average of the two closely agreeing recalculated analyses of miser-

TABLE 2. AVERAGE OF TWO RECALCULATED ANALYSES OF MISERITE, WITH RATIOS

Average of recalculated analyses		Ratios		Calculated for $K_2O \cdot 8CaO \cdot 10SiO_2 \cdot 3H_2O$
SiO_2	49.25	0.8333	10.04 or 10	50.16
TiO_2	0.43			—
Al_2O_3	0.81			—
Fe_2O_3	0.42	—		—
CaO	36.53	0.6561	7.97 or 8	37.46
FeO	0.06			—
MnO	0.28			—
Na_2O	0.96	0.0802	0.98 or 1	—
K_2O	6.09			7.87
H_2O	5.17	0.2869	3.51 or 3	4.51
	100.00			100.00

ite given in Table 1, with the derived ratios, is shown below in Table 2.

Even if 5 per cent of wollastonite (a maximum) is considered to be present in the sample and deducted, the ratios become:

SiO_2	10.12 or 10×1.01
CaO	7.85 or 8×0.98
$(K, Na)_2O$	1.03 or 1×1.03
H_2O	3.66 or 3×1.22

resulting in the same formula.

The derived formula of miserite is $K_2O \cdot 8CaO \cdot 10SiO_2 \cdot 3H_2O$ or $KCa_4Si_5O_{13}(OH)_3$. It is believed that the small excess of water found in this fibrous mineral is extraneous, owing possibly to the many washing operations to which the analyzed sample was subjected.

Finding that the dominant alkali present was potassium and not sodium was so surprising that I asked two of my colleagues in the Geological Survey to determine the alkalies in the same sample for confirmation of my results. The results obtained are as follows:

	R. E. Stevens	M. Fleischer	W. T. Schaller	Average
Na_2O	0.94	0.84	0.96	0.91
K_2O	6.33	6.38	6.39	6.37

A spectrographic determination of lithium by K, J. Murata indicated that lithium was of the order of 0.01 per cent.

The gross errors in Brackett's alkali determinations recall similar gross errors in the reported values for Na_2O and K_2O in several micas by different authors, as already noted by Schaller and Stevens (1941), pp. 542–543). To emphasize the magnitude of these errors, the values may be repeated here.

Mineral	*Reported*		*Correct*	
	Na_2O	K_2O	Na_2O	K_2O
Polylithionite	7.6	5.4	0.5	11.1
Hallerite	7.6	3.1	3.0	8.9
Muscovite	5.3	2.7	1.8	7.6
Euphyllite	3.6	4.6	5.6	1.7
Miserite	4.4	0.9	0.9	6.4

For polylithionite and euphyllite the presence of lithium may have contributed to the errors but for the other minerals the errors are analytical. A possible explanation of the gross errors in Brackett's figures for the alkalies may be that in transcribing the results from his laboratory notebook, he inadvertently transposed the figures for Na_2O and K_2O.

I was much intrigued by the formula obtained from Brackett's analysis, as it agreed closely with that of a compound intermediate between the formulas of wollastonite and of pectolite, if the latter be considered as derived from that of wollastonite by the double substitution of one atom of sodium for one atom of calcium with compensatory substitution of one hydroxyl for one oxygen, as follows:

Wollastonite	Ca_6	Si_6	O_{18}
Intermediate	$NaCa_5$	Si_6	$O_{17}(OH)$
Pectolite	Na_2Ca_4	Si_6	$O_{16}(OH)_2$

A comparison (Table 3) of Brackett's analysis with the calculated composition of the intermediate formula shows close agreement.

TABLE 3. COMPARISON OF BRACKETT'S ANALYSIS WITH THE CALCULATED COMPOSITION OF THE INTERMEDIATE COMPOUND $NaCa_5Si_6O_{17}(OH)$

	Brackett's analysis		Intermediate formula $NaCa_5Si_6O_{17}(OH)$
SiO_2	50.96		52.93
CaO	36.72	40.38	41.19
FeO	1.69		—
MnO	1.40		—
MgO	0.57		—
Na_2O	4.41		4.55
K_2O	0.90		—
H_2O	2.74		1.33
	99.39		100.00

If the composition of miserite was that represented by Brackett's analysis, that is, intermediate between wollastonite and pectolite, then the specific gravity and indices of refraction likewise should be intermediate but they were found not to be so, as shown:

	Specific Gravity	*Indices of Refraction*
Wollastonite	2.91	1.620 to 1.634
Miserite	2.84	1.58 to 1.59
Pectolite	2.86	1.595 to 1.632

Miserite is not an intermediate compound between wollastonite and pectolite, either in composition or in its other properties, although Brackett's analysis suggested such a relation.

DEHYDRATION

Dehydration determinations, made by Dr. P. G. Nutting, on miserite and, for comparison, on xonotlite from Puebla, Mexico, show considerable differences. Xonotlite continuously loses a few tenths of a percent of water from 110°C to about 700°, when most of its total water (about 3 per cent) is given off. Total loss to 655° is 0.80 per cent; to 940° it is 3.30 per cent. Miserite, on the other hand, loses no water until about 250° is reached. From here on to 980° there is a continuous loss. Total loss to about 800° is 2 per cent. Above 800° it loses an additional 3 per cent. When plotted the two dehydration curves are not similar, as shown in Fig. 1.

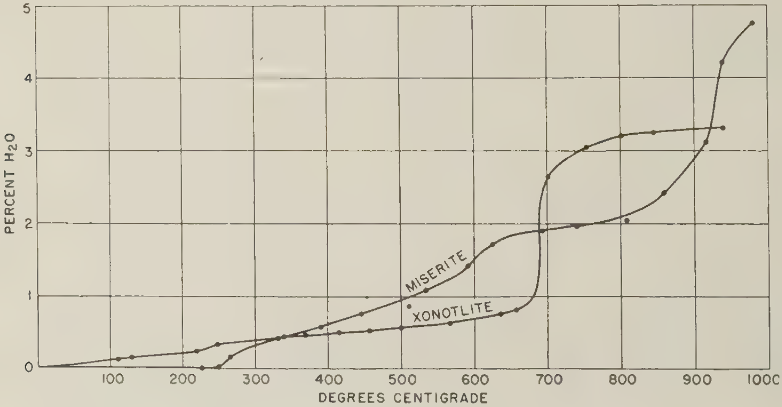

FIG. 1. Comparison of loss of water in miserite and xonotlite.

The loss of water by dehydration of miserite and xonotlite at increasing temperatures is given in Table 4.

TABLE 4. TOTAL LOSS OF WATER OF MISERITE AND XONOTLITE AT DIFFERENT TEMPERATURES

Temp. ° C.	Total loss		Temp. ° C.	Total loss	
	Miserite	Xonotlite		Miserite	Xonotlite
110	—	0.12	565	—	0.63
128	—	0.14	592	1.41	—
218	—	0.21	625	1.72	—
227	0.00	—	633	—	0.74
250	0.04	0.33	655	—	0.80
265	0.16	—	695	1.89	—
333	—	0.43	702	—	2.65
340	0.44	—	740	1.96	—
370	—	0.48	755	—	3.03
390	0.57	—	802	—	3.20
415	—	0.51	807	2.06	—
445	0.78	—	845	—	3.24
455	—	0.53	860	2.41	—
500	—	0.58	917	3.11	—
512	0.84	—	940	4.23	3.30
535	1.09	—	980	4.77	—

X-RAY PATTERNS

That miserite is not a xonotlite has already been suggested by the differences in the dehydration curves of the two minerals. The evidence is further substantiated by a comparison of the x-ray diffraction patterns. The pattern of miserite is different from the patterns of wollastonite, pectolite, and xonotlite. Those of wollastonite and of pectolite are closer to one another than those of any other pair of these minerals. The great similarity between the morphology of wollastonite and pectolite (both triclinic) has been shown by Peacock (1935, p. 110). Of these four minerals the greatest difference in x-ray patterns is shown by miserite. In no sense is the pattern of miserite intermediate between those of wollastonite and pectolite. The patterns of miserite and of yuksporite also are different. Some analyses of yuksporite, a mineral related to pectolite, show high K_2O. In an abstract (*Mineral. Abs.* (London), vol. 3, p. 112, 1926) of a paper on pectolite, yuksporite is said to be "related to Na-xonotlite."

The formula of miserite ($KCa_4Si_5O_{13}(OH)_3$) cannot be derived directly from that of wollastonite but it can be derived from that of xonotlite ($Ca_5Si_5O_{14}(OH)_2$) by substituting one atom of potassium for one atom of

Table 5. *d*-Spacings of Miserite and Related Minerals

Miserite	Wollastonite			Ignited xonotlite	Pectolite		Xonotlite			Pseudowollastonite	Ignited miserite
1	2a	2b	2c	3	4a	4b	5a	5b	5c	6	7
16.0 s	7.7 mw			7.8 w	7.8 w 7.0 mw	7.7 2 6.9 2	7.0 s	7.06 7	7.10 6		
					5.5 mw	5.4 2				5.7 w	
							4.27 m	4.24 9	4.22 8		
	3.85 ms	3.86 4	3.84 s	3.87 m	3.90 ms	3.85 6		3.89 9			3.85 mw
	3.52 s	3.51 4	3.50 s	3.53 ms	3.51 m	3.58 5	3.63 s	3.62 3	3.61 10		3.53 m
										3.44 mw	
	3.33 vs	3.30 8	3.30 s	3.33 s	3.32 s 3.28 s	3.28 7	3.25 vs	3.23 9	3.24 8	3.24 vs	3.33 m 3.24 vs
3.15 vs 3.07 m	3.09 ms	3.08 4 2.97 10	3.07 ms 2.97 vs	3.11 m 3.00 vs	3.09 vs	3.08 9	3.09 ms	3.06 10	3.07 8		3.10 mw 2.99 s
2.94 s						2.89 10					
2.82 mw 2.78 mw							2.82 m	2.82 9	2.82 6	2.81 s	2.81 s
					2.73 m	2.72 6					
2.68 m							2.70 ms	2.69 7	2.68 6		
		2.54 4 2.46 4	2.54 mw 2.48 m	2.49 mw	2.59 ms	2.58 6		2.50 7	2.49 6		
	2.30 m	2.33 6	2.35 ms	2.36 mw 2.31 mw	2.30 s	2.31 7					
		2.27 5 2.17 7	2.29 ms 2.18 ms	2.20 mw		2.28 7 2.15 7					
2.10 mw		2.01 4	2.01 vw				2.04 mw	2.03 9	2.03 8	2.00 m 1.981 s	1.982 s
							1.945 m	1.938 9 1.827 7	1.940 8 1.826 6		
	1.759 mw	1.821 7 1.751 4	1.826 mw 1.753 m	1.768 w 1.734 mw	1.749 mw	1.77 7					
		1.710 8	1.716 s			1.705 9 1.698 7	1.712 mw	1.702 7	1.705 6		
1.672 m		1.599 7	1.602 s								
						1.539 7		1.512 7			
		1.473 7 1.451 6	1.478 ms 1.453 ms		1.488 mw	1.477 6					

References to Spacings Shown in Table 5

Miserite:
1. Analyzed sample. Arkansas.
Wollastonite:
2*a*. U. S. Geological Survey film 161. With strong preferred orientation. Kern County, California.
2*b*. A.S.T.M. Average.
2*c*. Neues Jahrb., 1948 A, p. 58.
Ignited xonotlite:
3. Puebla, Mexico. From same sample as xonotlite 5*a*.
Pectolite:
4*a*. U. S. Geological Survey film 164. With strong preferred orientation. New Jersey.
4*b*. A.S.T.M. Average. New Jersey.
Xonotlite:
5*a*. U. S. Geological Survey film 494. Analyzed sample. Puebla, Mexico.
5*b*. A.S.T.M. Puebla, Mexico.
5*c*. A.S.T.M. Isle Royal, Michigan.
Pseudowollastonite (synthetic):
6. From Geophysical Laboratory, Carnegie Institution.
Ignited miserite:
7. From same sample as miserite 1. Arkansas.

calcium, in the same way that the formula of pectolite can be derived from that of wollastonite:

Xonotlite	$Ca_5Si_5O_{14}(OH)_2$
Miserite	$KCa_4Si_5O_{13}(OH)_3$

But the differences in *x*-ray patterns of miserite and xonotlite are much greater than those of wollsatonite and pectolite.

Ignited xonotlite gives a pattern similar to that of wollastonite. Ignited miserite gives a different pattern, and may be a two-phase product; if it is, the similarity between one of these phases and pseudowollastonite appears to be close.

The patterns of these minerals were taken by J. M. Axelrod, who also submitted the measurements of the *d* spacings (filtered Cu radiation) given in Table 5.

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ROCK-MAGMA AND ROCK-SPECIES

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ABSTRACT

A discussion of two fundamental petrologic terms. The word *magma* has lost its former physical significance and become a mere sack-name for any hot silicate fluid. This has led to misunderstanding of the actual physical state of the igneous fluid at the moment of intrusion or extrusion. The extravagant multiplication of rock names leads the writer to offer a definition of *rock-species* and to advocate the use of phase-petrology.

1. ROCK-MAGMA

The word *magma* has a long history. Apart from its ancient Greek connotation (fruit pulp?), it was used for several centuries in pharmacy with the meaning of a paste of solid and liquid matter such as *magnesia magma* or milk of magnesia. From pharmacy the word passed into chemistry. It was included in Thorpe's *Dictionary of Chemistry* as late as the edition of 1922, with the definition "a paste or semifluid mixture." It was therefore a purely physical term, denoting a state of matter without regard to its origin or chemical composition.

About the middle of the 19th century this word began to be used by geologists in place of the older expressions "surface lava" and "subterraneous lava," which had been used from the time of James Hutton. The borrowed word was employed at first with strict regard to its physical connotation, but in the course of time this connotation has been forgotten and to most geologists "magma" means only "fluid rock substance," regardless of its physical condition. This is an unfortunate state of affairs, for it leaves us without any special term to indicate a partly crystalline lava, and what is worse, it opens the door to misunderstanding of the properties of intrusive igneous matter.

The first geologist to realize that lava is typically not a homogeneous liquid but a suspension of solid particles in a liquid medium was George Poulett Scrope, English volcanologist. In the first edition of his *Considerations on Volcanos* (London, 1825), Scrope wrote that

> "The liquidity of lava, even at the moment of its protrusion from the orifice of a volcano, is extremely imperfect and of a very peculiar character. There is no reason for supposing it to be a homogeneous liquid like water or the metals and earths in a state of complete fusion. On the contrary, every circumstance in its conduct and composition shows its analogy to those compound liquids, such as mud, paste, milk, blood, honey, etc., which consist of solid particles deriving a certain freedom of motion amongst one another from their intimate admixture, in greater or less proportion, with one or more perfect fluids which act as their vehicle."

This statement is no less than a claim that lava is a magma in the physical sense, although the word magma was not used. But in the second edition of this book, which appeared under a slightly different title (*Volcanos, the Character of Their Phenomena*, etc.) in 1862 we find the terser statement that

> "lava is not a homogeneous liquid such as any melted or completely fused substance, but a *magma*, or composed of crystalline or granular particles to which a certain mobility is given by an interstitial fluid."

This is the conclusion reached by an experienced volcanologist from his own observations on Mount Vesuvius in one of its most active periods, and on other Mediterranean volcanos. Scrope's opinion was accepted in full by the master petrographer, H. Rosenbusch, who in the latest edition (1908) of his *Mikroskopische Physiographie* wrote

> "Eine Lava ist im Momente ihres Ergusses nicht eine reine schmelzflüssige Lösung, sondern eine schmelzflüssige Lösung in welcher grössere oder kleinere Mengen ausgeschiedener Mineralien in idiomorpher Begrenzung schwimmen." (A lava at the moment of its extrusion is not a pure melt but a melt in which larger or smaller numbers of idiomorphic crystals are already swimming).

Among recent writers who seem to have overlooked the historic meaning of the word magma are F. F. Grout, who has said that "the term magma primarily denotes a liquid," and P. Niggli, who defines magma as "eine glutheisse molekulare Lösung." But both writers admit that some crystals *may* be present in magma. A. K. Wells (1949), says only that "lava is molten rock material—magma—poured out through a vent or fissure." The most complete repudiation of the historic meaning of magma was pronounced by H. H. Read in 1947. In opening a symposium on the origin of granite (Ottawa, 1947) Read said:

"The igneous rocks are those produced by consolidation of *magma*, which is completely fluid rock substance." Since gas is excluded from consideration, the expression "completely fluid" can only mean completely liquid. Before we accept a definition so flatly opposed to the teaching of Scrope and of Rosenbusch, let us see what evidence we possess regarding the physical condition of the "stuff of which the igneous rocks are born" (Bowen's phrase).

(1) Various observers have recorded visible crystals in flowing lava, notably leucite and augite, on Vesuvius, and olivine, on Hawaii. Scrope himself pointed out that "the scoriform crust which rapidly congeals on the surface of a current by exposure to the open air, is found to possess the same crystalline texture, and to contain the same proportion of larger or minute crystals, as the innermost parts of the current" (1825, p. 21). This observation would be difficult to explain unless the larger crystals were already in existence before the crust was formed.

(2) About many volcanos of the more explosive type there are deposits of "crystal tuff," consisting largely of idiomorphic crystals, often of moderate size, of various minerals (especially plagioclase, augite, leucite, even hornblende and biotite) which must have been set free by the explosive dispersion of the liquid matrix in which

they grew. The summit of Mt. Vesuvius at the present time is strewn with myriads of perfect little augite crystals. Scrope recorded that after the eruption of Etna in 1669, "aeriform explosions succeeded from the same orifice, and continued to be discharged with violence during fourteen days. The fragmentary matters eructed by them produced the large cone called Monte Rosso, near Nicolosi, and covered a circuit of about two miles radius with a deep deposit of black sand containing innumerable separate crystals of augite" (1825, p. 152).

(3) Porphyritic texture is characteristic of lavas in general, and it has always been held to indicate two periods of crystallization, the large crystals (insets) having been formed before the lava reached the surface. In some basalts it has been observed that the augite insets carry thin mantles of pigeonite, while the groundmass in which the insets lie contains pigeonite as the only pyroxene. The explanation must be that the augite crystals were formed before the pigeonite and under different physical conditions, the change of conditions coinciding with the extrusion of the lava. (This observation, recorded long ago by Rosenbusch, has lately been repeated by G. A. Macdonald in Hawaii.)

(4) Even the most glassy of lavas, when examined in large bodies, will often be found to contain scattered crystals. Scrope wrote, of obsidian flows, that "whatever crystals appear embedded in these rocks, rendering them porphyritic, are clearly not of subsequent formation, but existed in them prior to the emission of the lavas, as is proved by their broken and half-fused state" (1825, p. 117). To the writer, the most striking demonstration of this fact is given by Pele's Hair, which consists of threads of perfectly transparent basaltic glass formed by the drawing-out and almost instantaneous chilling of droplets of intensely hot lava, shot into the air by the lava fountains of Kilauea. Here and there on the threads one may see tiny beads or swellings, and the microscope often reveals idiomorphic crystals or rounded grains of olivine, less commonly crystals of magnetite, in these beads. T. A. Jaggar showed that the temperature of the lava in the fountains may reach 1350°, so even at that excessively high temperature the lava contained a few crystals.

In view of this evidence, it will surely be conceded that some lava (certainly), much lava (probably), and even all lava (possibly) is already a suspension of crystals in liquid—a magma in the original sense of the term—at the moment of its extrusion. Perhaps this is not, in itself, a matter of great importance, but it raises a very important question, namely: what is the physical condition of deep intrusive matter, the "subterraneous lava" of Hutton? Is it hotter or cooler than surface lava? Is it a homogeneous liquid or is it a magma?

Many geologists, impressed by the geothermal gradient in sedimentary rocks, have assumed that active lava too must grow hotter downwards. But although sedimentary rocks, starting at atmospheric temperature, become hotter downwards, it does not in the least follow that lava with a surface temperature of 1000° must also grow hotter downwards. At Kilauea, Jaggar was able to show that the surface of the lava received much heat from burning gases, and that the temperature of the lava a few feet beneath the surface was sensibly lower than that of the surface layer. It seems impossible that basalt at 1000° can grow hotter downwards, because the common rocks of the earth-crust begin to soften and melt at temperatures of that order. Sosman and Merwin (1913) found that the arkose that underlies the Palisades diabase was more than half melted in 75 minutes at 1100–1150°; and Greig, Shepherd and Merwin observed (1933) that a specimen of powdered granite became half liquid in one week at 800°. It must follow that a deep-seated reservoir of basalt, at a temperature of more than 1000°, would inevitable melt and dissolve its walls until the contents of the reservoir no longer had the composition of basalt. Yet the enormous outpourings of basalt in the Deccan, the Columbia River region, and the Parana basin show no significant change in composition from the earliest flows to the latest.

Great dikes, sills, laccoliths and batholiths give no evidence of very high temperature. The 1000-foot sill of the Palisades did not even begin to melt the invaded Triassic arkose. Many diabase dikes in South Africa contain fragments of granite which have not been vitrified or appreciably corroded, although similar inclusions in surface basalt flows are often half-melted (e.g. Parícutin). The evidence of the peridotite-serpentine dikes which cut coal seams in Pennsylvania, and enclose fragments of coal, is very significant. From the comparison of these enclosures with pieces of the same coal heated in the coking furnace, R. B. Sosman (1938) found it possible to assert that the intrusive fluid can not have been much hotter than 520°. Sosman assumes this fluid to have been a mush of olivine crystals in a watery matrix. One is reminded of the remarkable olivine-bearing rock which was called "maimechite" by Moör and Sheinmann (1946). This is a strongly banded dike rock, some of the bands being rich in olivine crystals while other bands are largely and in some cases completely glassy. The discoverers claim with reason that the intrusive matter was "a viscous mass containing a considerable quantity of olivine crystals."

Useful evidence may also be obtained from the study of texture. In some dikes there is a marked orientation of tabular feldspar insets parallel to the walls of the dike, indicating that these insets were present in the magma at the time of intrusion. H. Winkler (1949) has pointed out that in the Cleveland dike of the north of England the plagioclase insets are "as well developed and of the same size" in the chilled margin as in the centre of the dike; from which it must follow that these crystals were present in the magma when it was injected. A similar observation was recorded by R. A. Daly (1912) in the Rock Creek porphyry of British Columbia, the insets of anorthoclase and augite having the same dimensions in the chilled margin as in the central parts of this great dike.

In the case of really deep intrusions we have to rely on the "geological thermometer" method of study. The entire absence of any evidence of the quartz-tridymite transformation, among deep-seated intrusions and their contact rocks, limits the maximum temperature of intrusion to little more than 870°. The presence of wollastonite at some granite-limestone contacts and its absence from others seems to restrict the temperature of emplacement of granite to the region of 500–550°. A well-developed porphyritic texture is characteristic of very many granites, and it is sometimes possible to demonstrate that the history of the large feldspar insets has been different from that of the groundmass feldspar (Shand, 1949).

We seem to be justified in concluding that *all igneous matter, within such levels of the earth-crust as are exposed to observation, was at the time of its emplacement a magma in the physical sense and not, or only in rare instances, a homogeneous liquid.*

It remains to indicate the importance of the conclusion we have just reached. In the first place, it has been assumed in certain theories of the earth that deep-seated igneous rocks have been generated by superheated liquids that are able to "melt and stope" their way up through the crust. Now a homogeneous liquid may be superheated to any degree short of vaporization or decomposition, but a mush of crystals in their mother-liquid cannot be superheated. If deep rock-magma is such a mush, then there can be no such thing as a superheated liquid in the higher levels of the lithosphere. Yet many outstanding petrologists have written about "superheated magma."

Though it cannot be superheated, a rock-magma may nevertheless be fluid through a considerable range of temperature. A quantitative illustration may be found in Bowen's study of the melting phenomena of the plagioclase feldspars. A mixture of the composition Ab_1An_1 is completely liquid at 1450°. With moderately slow cooling it will be half crystalline at 1386°, three-quarters crystalline at 1282°, seven-eights crystalline at 1164°; thus the possible range of temperature, for this anhydrous feldspar magma, is of the order of 300°. The liquid phase in this system is rather viscous so one cannot be sure at what temperature fluidity will cease, but with a less viscous fluid it is only necessary to have enough liquid present to act as a lubricant between the crystals. It does not seem fantastic, then, to suggest that a granitic magma, saturated with water, may be fluid through a range of temperature of something like 800° down to 500° or less. Thus many problems of igneous intrusion are capable of solution without the assumption of superheat.

A second instance of faulty reasoning, resulting from forgetfulness of the physical constitution of a magma, may be found in the attempts that have been made from time to time to compare the density of a particular kind of rock with that of a particular magma, in order to decide whether enclosed blocks of the rock in question should sink or swim in the magma. Since the density of natural rock-magma cannot be determined directly, the practice has been to melt a rock of the appropriate composition, chill it to the state of glass, and measure the density of the glass. But such a glass is free from crystals, while natural rock-magma, as we have seen, may contain anything from a few crystals up to perhaps 90% of crystals. Thus the density of a granitic magma (if it is permissible to speak of the "density" of a mixture of solid and liquid) must lie between that of average granite (2.66, according to Daly) and that of rhyolitic glass (2.36, according to Tilley); and any computation based on the assumption that granitic magma has a fixed density, can have little value. In any case, if the magma is far advanced in crystallization, the factor which determines whether an inclusion shall sink or float may be mere mechanical obstruction by the mass of interlocking crystals.

Surely we have said enough to show that the loose use of "magma" to mean merely "rock-forming fluid," without consideration of its physical state, is to be deplored. Yet, as an anonymous Briton has said:

> "It is a cursed thing, but true, that if a sufficient number of people speak wrongly a sufficient number of times, their wrong speech becomes right."

So there is little reason to expect that petrologists in general will renounce their slackness in the use of "magma." The most one can hope for is that they will bear in mind that although rock-magma is fluid rock substance, only a minor part of it need be liquid. The properties of the magma are largely determined by the presence of crystals in it and by the proportion of crystals to liquid.

2. Rock Species

If the term "magma" has been misused by petrologists, what shall we say about "species"? We have at present some 700 rock names of specific character, yet we have no accepted definition of what constitutes a rock species. Exactly 50 years ago C. R. van Hise wrote these words:

> "The method of petrographers in giving names, so far as any method is discoverable, is to give an independent name to each rock which is slightly different from any previous rock found, without regard to any definite plan."

These words are as true today as they were 50 years ago. Five new names have lately been added to the list of pyroxenites and peridotites, which already included 66 names. The leucitic lavas of Western Australia are strikingly similar to those of the Leucite Hills in Wyoming, for which Cross introduced three specific names, yet the discoverers of the Australian lavas felt it necessary to introduce four entirely new names. Where is this practice to end? If the same procedure were followed in the classification of animals, then every mongrel dog and every alley cat would be a new species on the ground that it is "slightly different" from the last one seen!

The difficulty of defining a species is not confined to petrology. An eminent zoologist has laid it down that "a species is what a competent systematist considers to be a species." But which of us is a competent systematist, and how do we know? Of the thirty or more classifications of rocks that have been proposed in the past 50 years, not one commands general acceptance. In the absence of agreement among petrologists, the writer offers his own definition of a rock species, as follows:

(A) A rock species is a group of rocks formed under similar physical and chemical conditions.

(B) Two rocks belong to the same species if both contain the same dominant minerals (with some latitude for ionic substitution) and if no phase in one rock is incompatible, either chemically or physically, with any phase in the other rock.

Examples of physically incompatible minerals are the following pairs: tridymite-microcline, tridymite-muscovite, sanidine-muscovite, leucite-microcline, leucite-muscovite (both minerals in each pair being primary). Examples of chemically incompatible minerals are the following pairs: quartz-olivine, quartz-feldspathoid, hypersthene-feldspathoid, muscovite-aegirine, anorthite-aegirine (both minerals in each pair being primary).

Both chemical and physical incompatibility are indicated, in magmatic rocks, by such associations as pyroxene-tourmaline and olivine-tourmaline.

(C) Differences in the relative proportions of phases do not yield new species but only *facies*. These should not receive distinctive names but only adjectival prefixes, as leucocratic syenite, melanic gabbro, etc.

(D) Differences in texture, except that between holocrystalline and hypocrystalline (or plutonic and volcanic) rocks, do not yield new species but only *varieties* which should be distinguished by the appropriate adjectives, porphyritic, poikilitic, pegmatitic, granophyric, etc.

Reconsideration of the existing 700 "specific" names, in the light of these propositions, would result in the immediate elimination of more than half of the names in question and perhaps, after fuller consideration, of half of the remainder.

The admission that petrology is a province of physical chemistry should be enough to condemn out of hand not only the old "natural history" type of classification but also the type that depends entirely on bulk chemical analysis. Several attempts at a classification of the latter kind were made in the early days of petrology, before the microscope came into use; but Justus Roth, who compiled the first large collection of rock analyses (1861), rejected the idea that any progress in understanding the genesis of rocks could be made by bulk analysis alone. Yet the idea was revived by Osann in 1899, by the authors of the Norm system in 1903, and by Niggli in 1923. Undoubtedly a systematic tabulation of rock analyses has many uses—few people can have made more use of Washington's *Tables* than the writer—but a tabulation of analyses is a very different thing from a classification of rocks. A chemical system consists of components, which enter into the system, and phases, which crystallize out. To discuss a rock in terms of components, without regard to phases, is just as unsatisfactory as it would be to discuss language in terms of sound without regard to sense.

As a step toward the petrology of the future, the writer (1942) has used the method of "phase petrology" (in full, *critical phase petrology*). This method depends on the recognition of certain phases, each of which has some special significance for the interpretation of the cooling history of a rock complex. Of course every mineral has some significance for this purpose, but some have more than others. From the early days of petrology an exaggerated importance has been attached to the feldspars; yet the feldspars tell us relatively little about the cooling history of a magma because they are very stable minerals which occur in almost all possible associations and within a wide range of temperature. Certain other minerals have a narrower range of stability and consequently afford more information about the physical and chemical conditions that prevailed and the changes that took place during the crystallization of a magma. These are critical phases, and the following examples may serve to indicate their use.

The presence of tridymite or leucite, and especially the association of leucite with a siliceous glass, indicates an unusually high temperature at the time of their formation.

The presence of microcline, low-quartz, tourmaline, aegirine, or muscovite indicates a temperature about the lower limit of the magmatic range.

The presence of feldspathoids, melilite, olivine, indicates a deficiency of SiO_2 in the system.

The presence of melilite, melanite, scapolite, cancrinite, probably indicates contamination of the system with limestone.

The presence of cordierite, andalusite, corundum, common garnet, and perhaps hypersthene, indicates contamination with alumina or aluminum silicate.

The presence of alkali-pyroxene or alkali-amphibole indicates an excess of sodium ions over aluminum ions in the system.

The presence of amphibole, mica, tourmaline, epidote, indicates reduced temperature and increased concentration of H_2O.

In the Cortlandt complex of New York State the most abundant minerals are plagioclase, augite, and hypersthene, but these minerals tell little of the history of the complex because they occur almost everywhere in it. On the other hand, prismatic hornblende, poikilitic hornblende, olivine, and spinel are restricted to certain regions in the complex, and a map showing the distribution of these critical phases conforms better to the structure, and gives a clearer picture of the history, of the complex than the older type of map based on the concept of natural rock species.

The method of critical phases is used in metamorphic petrology. A geologist studying a region of metamorphic rocks does not set out with a rigid concept of "natural species," but he traces the distribution of critical phases which indicate changes in the intensity and character of metamorphism. The classic example is the mapping of chlorite, biotite, garnet, staurolite, kyanite and sillimanite zones by G. Barrow in the Scottish Highlands.

Igneous petrology must undergo drastic changes. The period of collecting strange-looking specimens and inventing new names for them is over, one hopes. Our new duty is to study the accumulated data in the light of physical chemistry, to distinguish the important from the incidental, and to remember that the purpose of scientific classification is not to make a catalog of oddities but to help in the discovery and expression of natural laws. This requires the abandonment of old ideas about natural rock species, petrological affinity, spontaneous differentiation, and the like, and an increasing reliance on physical chemistry, until it is realized by everybody that petrology is the very root and trunk of geochemistry.

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THE NABABEEP NEAR WEST TUNGSTEN MINE, SOUTH AFRICA

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ABSTRACT

A number of ferberite-wolframite deposits in the Namaqualand copper district are restricted to several bands of sedimentary schist separated by paragneiss. These Archaean formations are intruded by Pre-Cambrian granite which was the source of the tungsten-bearing solutions. This paper deals briefly with the main features that influenced mineralization in the Nababeep Near West mine and with the mineralogy of the ore.

INTRODUCTION

Nababeep is situated about 300 miles north of Cape Town in the well dissected escarpment belt of Little Namaqualand separating the interior plateau from the Atlantic coastal plain. The copper deposits of this district have been mined since 1850 and with the discovery of the spectacular alluvial diamond deposits at Alexander Bay in 1927 attention was drawn to the rarer minerals, e.g., beryl, scheelite, tantalite, euxenite, bismuth, spodumene, in the pegmatite area 50 miles north of Nababeep. The revival of the tungsten market in 1933 gave further impetus to the search for scheelite which was subsequently found in numerous small vein deposits in metamorphosed basic eruptives along the Orange river over a distance of 200 miles.

Wolframite was discovered at Narrap in the copper district in 1938; the first successful mine was developed at Klein Kliphoog in 1941. In the same year six other mines started production, the largest of which was the Near West mine at Nababeep. At the end of World War II five had closed down while the Klein Kliphoog and Nababeep Near West mines suspended operations in 1949. The total production of the seven mines amounted to approximately 1,800 tons of concentrates, of which the Near West contributed 885 by treating 77,000 tons of ore averaging 0.92 per cent WO_3. No previous literature on these deposits has been published.

GENERAL GEOLOGY

The rocks of the copper district comprise a conformable succession of gneisses, granulites, quartzites and schists, correlated with the Archaean Kaaien series, intruded by granite of early Pre-Cambrian age. The Archaean formations are highly metamorphosed sedimentary beds largely transformed into rocks of granitic appearance, but retaining a more or less clear banding that represents the original stratification. Between Springbok and Nababeep the metasediments are folded into a major closed anticline whose limbs dip about 30° north and south while the two ends plunge down at about 10°. A minor elliptical dome appears immediately north of O'okiep. From Rooipoort to Steyerkraal in the north stretches a regional syncline with moderately inclined south limb and nearly vertical north limb. To the west these formations are capped unconformably by the early Cambrian Nama quartzite and shale. The subdivisions of the Kaaien series as established in the copper district are shown on the accompanying sketch map (Fig. 1).

Porphyritic Rietberg granite intrudes the gneisses and metasediments in the form of sills, dykes and irregular stocks regarded as offshoots from a buried batholith that caused the regional granitization of the ancient formations. Although the contact between granite and country rock is commonly sharp the latter is distorted only locally in minor degree; it is inferred that intrusion did not take place with strong directed force, but that the granite magma was largely generated by melting in situ accompanied by metasomatic transformation of the bedded formations. This mode of origin accounts for the development of transitional rock types that resemble granite but may have just escaped melting. Subsequently irregular noritoid dykes were injected into the metamorphic terrain and with these are associated all the important copper deposits. Oblique and transverse faults striking N.W., N. and N.E. are present in great numbers; their bearing on tungsten mineralization is locally important.

The Ore-Bearing Formations

All the more important tungsten deposits are located in the Wolfram schist, a marker stage about 400 feet thick, which separates the underlying Nababeep gneiss from the Concordia gneiss above. The lower part of this stage consists of varied granulites, siliceous hornfels, and gneiss; the upper 200 feet in the Nababeep area is dominantly banded sillimanite-quartz-biotite-garnet schist alternating with sills of leuco-granite, either intrusive or of metasomatic origin. In the Near West mine are a few lenticular interbeds of dark augite-amphibole schist containing variable amounts of quartz, labradorite-bytownite, biotite and garnet. While the larger part of the Wolfram stage appears to be highly metamorphosed shale the sillimanite-rich layers probably originated from kaolinic clay and the augite-amphibole schist from dolomitic marl. East of Nababeep the Wolfram stage is progressively changed to granulite, gneiss, and granitic rock so that only a few bands of schist, each a few feet thick are preserved at Klein Kliphoog and Concordia. South of the Springbok anticline the schist is very largely metamorphosed to quartz-feldspar granulite and only a few thin beds have been found at the Tweedam mine and in the Carolusberg area. In the east between these two belts

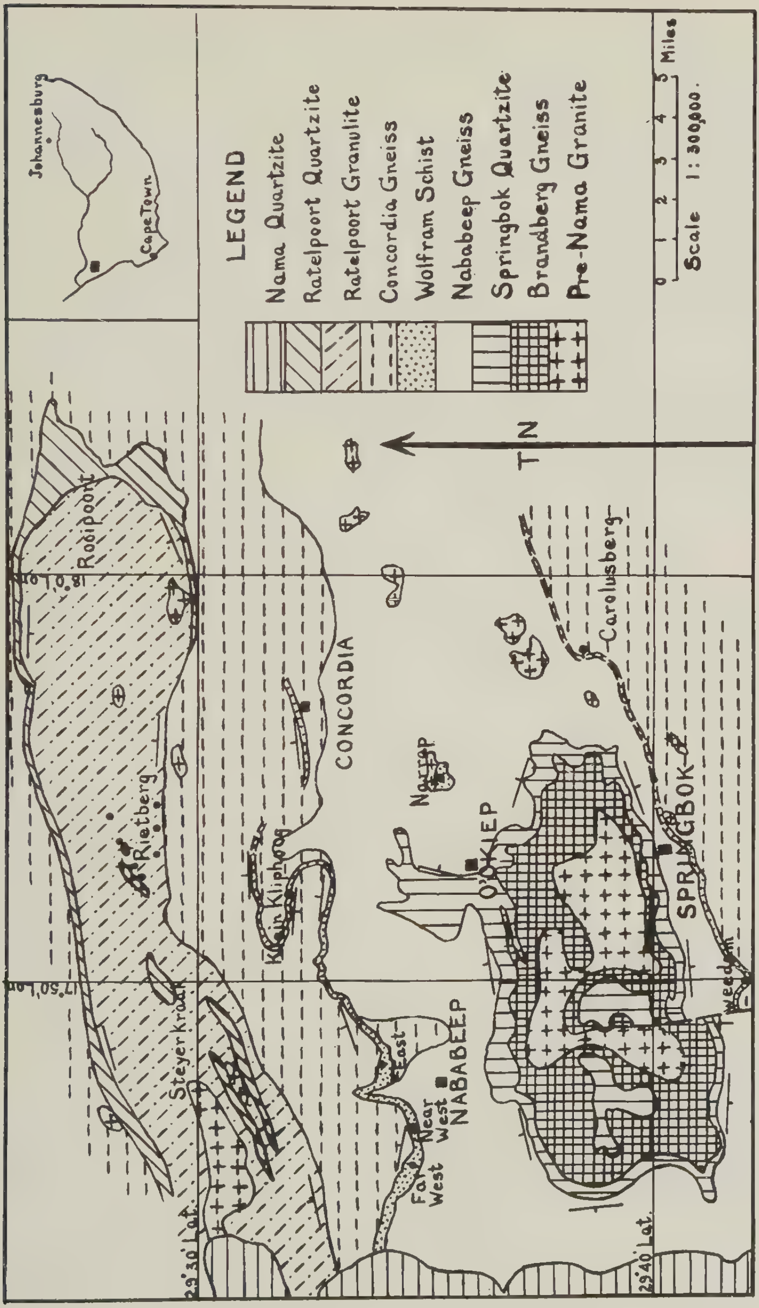

FIG. 1. Generalized plan showing regional geology and location of tungsten mines in the Namaqualand copper district.

of mineralization is the Narrap mine in a faulted outlier of the schist which here forms a band about 50 feet thick. There is no clear relation between the degree of metamorphism of the Wolfram stage and the proximity of intrusive granite, but the change is more complete where quartz-feldspar gneiss (Concordia type) appears concordantly within and below the schist. It is suggested that local fusion occurred at this horizon whence granite magma was injected over short distances and concurrent transformation of schist to metamorphic granite and granulite was intensified.

About 2,000 feet stratigraphically above the Wolfram stage are the Ratelpoort beds, now consisting mainly of red and gray granulite with occasional interbeds of gneiss, biotite-sillimanite schist and metaquartzite in which are located many wolfram prospects, only one of which—the Rietberg mine—produced a notable tonnage of concentrates.

And, lastly, two small showings of tungsten ore are found in the upper schist and quartzite of the Springbok stage, some 1,500 feet below the Wolfram schist. One of these yielded scheelite as well as wolframite.

STRUCTURE

The strike of the Wolfram schist swings from N 70° E at the east end of the mine through E-W, to N 70° W at the west end, the northward dip changing from 40° near the surface, through an average of 20° in the main section, to about 10° on the lower levels. The ore body, essentially conformable to the bedding of the host, lies therefore in a gentle synclinal cross-structure plunging to the north. East of the mine the schist swings sharply south due in part to drag along a major shear fault trending north while a group of anorthosite intrusions practically obliterate the Wolfram stage to the south; Concordia gneiss appears to have become mobilized here, replacing some of the schist. These disturbances have created a tight anticlinal cross-structure along which mineralization is interrupted, but farther east, where a second gentle cross-syncline is developed, payable reefs reappear.

Superimposed on the major synclinal structure in the mine are many minor rumples with a width of a few tens of feet and amplitude of 1–3 feet. While those in the upper part of the mine have probably developed by differential slip along the bedding during regional folding, some of the rumples in the deeper section are clearly associated with an oblique shear fault. Individual rumples generally die out within a distance of about 100 feet. Still smaller folds have been caused by injection of the vein quartz forming pods of ore.

There are only three major fractures in the mine. The first strikes N 40° W, dips steeply SW, and is exposed along the inclined shaft; pegmatitic granite forms an irregular dyke-sill intrusion in the distorted zone. The second shear fault, also containing pegmatitic granite, strikes N 45° E, dips nearly vertically and reaches from 9 level half-way up to 8 level before changing into a sharp fold that fades westward into a group

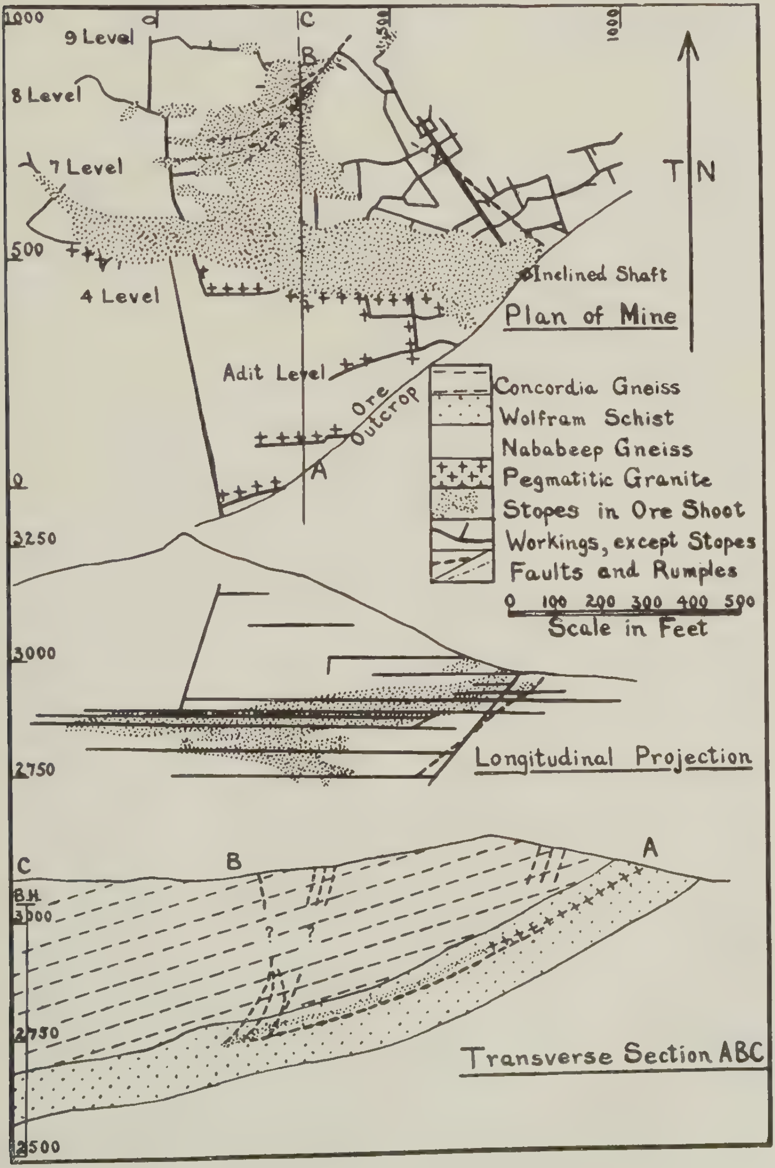

FIG. 2. Plan and sections of the Nababeep Near West tungsten mine.

of rumples; a gentle anticline also diverges from the fault on the south side. The third important fracture is a bedding thrust exposed throughout the mine about 5–20 feet below the ore zone. The schist has been ground up by the movement and is replaced by quartz and pyrite over a width up to 12 inches. Minor fractures striking due north and filled with quartz are most prominent below 7 level.

THE ORE SHOOT

In plan the ore body has a mushroom shape, the deeper part representing the stem and the upper part the curved top. Its length measures 1,000 feet, average breadth down dip 150 feet, with the stem portion extending down an additional 350 feet. The reef zone consists of a series of quartz veins parallel to the bedding through a stratigraphic distance of 10–20 feet in the upper half of the main band of schist. Individual veins swell out from an average width of 1½ feet to as much as 5 feet. The stoping width ranges from 3 feet to 12 feet.

By drawing contour plans of variations in dip and total thickness of minable reefs the following relations have been established:—(*a*) The richest and thickest reefs were localized in the upper east wing where dips are considerably steeper than the average and where irregular minor folding is evident. (*b*) In the central and western sections of the shoot a similar relation holds, though less clearly defined because changes in dip are less abrupt. (*c*) Below 7 level the richer ore is restricted to the medial part of the body as it rapidly narrows toward the distorted zone merging with the shear fault. (*d*) The anticlinal rumples, especially below 7 level, are loci of better grade reef, whereas synclinal rumples are commonly barren. Stoping toward the west was eventually all but confined to such rumples emerging from the shear fault.

Other factors in the localization of ore deserve mention:—(*a*) The degree of silicification of the schist and its conversion into hornfels prior to mineralization determined the extent to which the ore solutions could find their way into the host rock. The western boundary of the shoot is where the soft schist changes laterally to siliceous garnet hornfels which resulted from the formation of numerous sills of quartz-feldspar granite in that area. Similar silicification partly accounts for the lower eastern ore limit which is, however, an economic boundary for the reefs, though poorer, continue toward the shaft. (*b*) The bedding fault prepared the ground above for mineralization. (*c*) The NE shear fault on 9 level appears to have been the focus whence the ore solutions spread out upward along the bedding, guided in part by minor fractures striking north and by the anticlinal rumples already described. (*d*) The upper margin of the payable ore on 4 and 5 levels roughly coincides with the appearance of sericitized quartz-feldspar pegmatite merging with the quartz veins. The pegmatite constitutes low-grade ore.

Rock Alteration

Extensive *silicification* occurred during and after the formation of dykes and sills of pegmatitic granite. The sillimanite of such schist was largely altered to sericite which, with biotite, was progressively replaced by quartz, leaving only shadows of the digested minerals. Early *pyritization* was more restricted, though generally accompanying the silicification; the widespread late pyrite was deposited more by filling than by replacement of the host rock. *Pegmatitic recrystallization* of the schist alongside the quartz veins caused the formation of lenses of coarse-grained sillimanite, biotite and garnet up to 10 feet away from reef. Some of them contain ferberite crystals unaccompanied by quartz. *Sericitization and chloritization* on a moderate scale followed; biotite progressively lost its brown color to become a white to greenish nearly opaque substance, probably a mixture of chlorite, sericite, ilmenite and leucoxene. Pyrophyllite may accompany the sericite replacing sillimanite. Where ore entered along cracks in garnet and sillimanite fine chlorite was locally developed. Microcline was also sericitized.

The Quartz-Ferberite Veins

Form. There is a general change from thick, persistent veins, traceable over several hundred feet along strike, in the upper workings to lenses and pods of quartz separated by barren schist in the lower. The larger veins have structurally regular walls, whereas the schist wrapping around the pods is distorted and progressively replaced by quartz. The vein-forming solutions evidently entered the host rock under great pressure; once entry had been gained replacement of schist proceeded simultaneously from several bedding planes so that the final ore vein was much wider than that formed by the initial injection. The veinlets branching off the larger pods are commonly highly irregular, some showing forms resembling ptygmatic folds; as no crumpling is to be seen in the schist the snake-like vein pattern must have developed by alternate replacement along two or more cleavage planes.

Texture. Most of the quartz veins display ribbon texture, the banding being more or less parallel to the foliation of the schist. Where the latter has been distorted the ribbons show similar folding. Where the schist broke into thicker slabs of less regular shape relics in various stages of conversion into quartz cause a streaky to mottled texture between ribboning and slabby breccia. The veins hold up to 10 per cent garnet, biotite, chlorite and sillimanite, garnet tending to be concentrated along the contact. Segregations of ferberite and the accompanying sulphides commonly show a sub-parallel arrangement, especially near the margins of the veins, while locally individual ferberite crystals display such preferred orientation. The growth of the ore minerals was clearly influenced by the ribbon texture and by parallel cracks formed during metallization. In the upper levels the quartz merges with streaky sericitized pegmatite.

MINERALOGY:

Apatite, associated with pyrite, is an unusual constituent in some veins of fine-grained quartz and was probably derived from the schist in which it is a common accessory.

Bismuth and *bismuthinite* appear along cleavage traces in biotite, sericite, sillimanite and molybdenite and as corroding blebs and veinlets inside pyrite, ferberite, garnet and quartz. Together they constitute about 0.1 per cent of the ore.

Calcite is a rare associate of pyrite and chlorite.

Chalcopyrite in minor amount was deposited with early disseminated pyrite and again with the comparatively late bismuth minerals.

Chlorite with very low birefringence developed as a widespread alteration product during the replacement of schist by quartz and in microscopic amounts during metallization.

Ferberite forms tabular crystals, up to 2 inches in length, typically clustered in lenses parallel to the walls of the reef. Under the microscope ferberite may be seen as dust-size particles replacing sillimanite and biotite or as coarser grains intergrown with garnet and quartz. The mineral has $FeWO_4{:}MnWO_4 = 91{:}9$.

Fluorite is a rare constituent accompanying late scheelite.

Garnet grains form deep red aggregates up to 2 inches in diameter in the reefs. The average refractive index (1.793) and specific gravity (4.02) show little variation between garnet in the schist and that in the reef and roughly indicate a composition of almandine_{60} pyrope_{30} grossularite_{10}.

Jacobsite (?) or magnetite forms a grating in many ferberite crystals, especially near cracks, and also little blebs and veinlets distributed at random. Due to its very fine grain the mineral could not be identified with certainty. It may be a product of exsolution or of oxidation.

Microcline in curved crystals is largely sericitized in the pegmatitic reef of the upper levels.

Molybdenite, generally concentrated in streaks along the reef contact, forms corroded relics in ferberite and pyrite. Replacement of biotite is indicated by a semi-opaque fringe of molybdenite dust separating the solid crystals from the gangue.

Pyrite developed as an early fine-grained dissemination during wall rock silicification and crystallized again after deposition of ferberite and molybdenite to which it is adapted. Finally some pyrite filled numerous minute cracks in the reefs, developing euhedral forms where cavities existed. Such pyrite in the bedding fault suggests that this channel was open to the end of mineralization.

Quartz, the dominant gangue mineral, ranges from fine in partly replaced schist to very coarse and vitreous in the wider reefs, the color shading from bluish gray to grayish white. The interlocking grains have sutured boundaries and commonly show undulatory extinction, locally accompanied by clearer granulation. Late euhedral quartz is found in little vugs in the bedding fault.

Scheelite is sporadically present in small amount, fringing ferberite crystals and in occasional vugs where it is accompanied by fluorite. The augite-hornblende schist contains layers of scattered, tiny scheelite grains.

Sericite as tiny nests and veinlets in quartz and garnet is probably derived from the alteration of biotite and sillimanite. Partial chemical analysis by A. Knoetze established the presence of sufficient potash and soda to verify its identity, but some pyrophyllite may be mixed with it.

Sillimanite relics are found in the vein quartz in accessory amount, but the white prisms are most conspicuous in the pegmatized wall rock.

Sphalerite is contained in chalcopyrite as minute specks.

Zircon is an accessory doubtlessly inherited from the replaced schist.

The simplified paragenetic sequence is:

Pegmatitic Stage: Zircon and apatite (inherited), sillimanite, garnet, microcline, quartz.

Hydrothermal Stage:

(*a*) Quartz, sericite, chlorite, calcite, pyrite.

(*b*) Quartz, molybdenite, ferberite, scheelite (?), pyrite, bismuthinite, bismuth, chalcopyrite, sphalerite, scheelite-fluorite, jacobsite (?).

At the surface the ore zone has undergone moderate leaching and oxidation. *Bismutite* forms a greenish yellow impregnation particularly in sericitized pegmatite. Chalcopyrite has changed to sooty *chalcocite* and *covellite* besides causing green copper staining. Pyrite is oxidized to dark brown transported *limonite*, while ferberite shows all stages of alteration to reddish brown limonite with cellular boxwork. Sericitized microcline is largely changed to kaolinite.

ORE GENESIS

The regional setting of the tungsten mines of the Namaqualand copper district in several belts of schist and granulite separated by unmineralized paragneiss proves that the ore solutions entered and moved within the schist with far greater facility than in the gneisses. Where the upper part of the Wolfram stage is mainly schist the vein deposits are of the hypothermal type (e.g., Nababeep Far West and Near West, Narrap), but where the host rocks were thoroughly metamorphosed to granulite alternating with Concordia gneiss the veins are pegmatitic, ranging from pure quartz to pure microcline (Nababeep East, Klein Kliphoog, Rietberg, Tweedam). The type of ore was therefore conditioned by the grade of metamorphism, including granitization, which can only mean that the ore solutions were brought into circulation by the buried granite batholith that caused the widespread granitization and is indicated by the smaller intrusions exposed at the present surface. Analysis of the detailed structure in the Nababeep Near West mine shows how pre-existing folds and fractures have guided the ascending vein-forming fluid and localized the formation of ore shoots.

ON THE TEMPERATURE-PRESSURE-CONCENTRATION DIAGRAM FOR BINARY SYSTEMS IN WHICH THE ONLY CRYSTALLINE PHASES ARE THE PURE COMPONENTS

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ABSTRACT

An isometric drawing of the three dimensional temperature-pressure-concentration diagram for a hypothetical binary system in which there are critical end points and in which the only crystalline phases are the pure components is given and the significance of the fields is stated. This diagram is analogous to the three dimensional temperature-pressure-concentration diagram for a hypothetical binary system in which there is a continuous critical curve from the critical point of one component to the critical point of the other component, the only crystalline phases in the system being the pure components, that was published by H. W. Bakhuis Roozeboom in his treatise, *Die heterogenen Gleichgewichte.*

The relations of the fluid and crystalline phases during the formation of pegmatite dikes and hypogene ore deposits have been the subject of much discussion by geologists and geochemists in recent years. In these discussions two types of binary system have often served as starting points, namely, binary systems in which there is a continuous critical curve from the critical point of the first component to the critical point of the second component, and binary systems in which this critical curve has critical end points, both types of binary system having no intermediate crystalline compound and no solid solution.

A drawing and a photograph of a model of the three dimensional temperature-pressure-concentration diagram for binary systems in which there is a continuous critical curve from the critical point of the first component to the critical point of the second component and in which the only crystalline phases are the pure components were published by H. W. Bakhuis Roozeboom as Plates I and II of Teil 1 of Heft 2 of his treatise, Die heterogenen Gleichgewichte,[1] and the significance of the fields in the three dimensional diagram was explained by him on pages 125–129.

An analogous isometric drawing[2,3] of the three dimensional temperature-pressure-concentration diagram for binary systems in which there are critical end points and in which the only crystalline phases are the pure components is given here as Fig. 1. The significance of the various fields in this figure is as follows:

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[1] Friedrich Vieweg und Sohn, Braunschweig (1904).

[2] A perspective drawing of the three dimensional temperature-pressure-concentration diagram for a binary system with critical end points appeared in Boeke-Eitel's *Grundlagen der physikalisch-chemischen Petrographie*, 2te Aufl., Gebrüder Borntraeger, Berlin (1923), on page 346. In this drawing, however, the saturation surface of the fluid phase between the critical end points is not shown. The intersection of the surfaces representing the saturated fluids (liquids) with the upper plane of the parallelepiped is also not drawn.

[3] Niggli has given two series of sections of the temperature-pressure-concentration diagram for a binary system with critical end points, one series perpendicular to the pressure axis and the other perpendicular to the temperature axis (Das Magma und seine Produkte. I. Teil: *Physikalisch-chemische Grundlagen, Akademische Verlagsgesellschaft*, Leipzig (1937), pages 238–239). He also gave a perspective drawing of the three-phase surfaces, gas-liquid-solid, in the three dimensional diagram (*ibid.* p. 236), and an excellent discussion of equilibrium relations in systems of this type (as well as many other types).

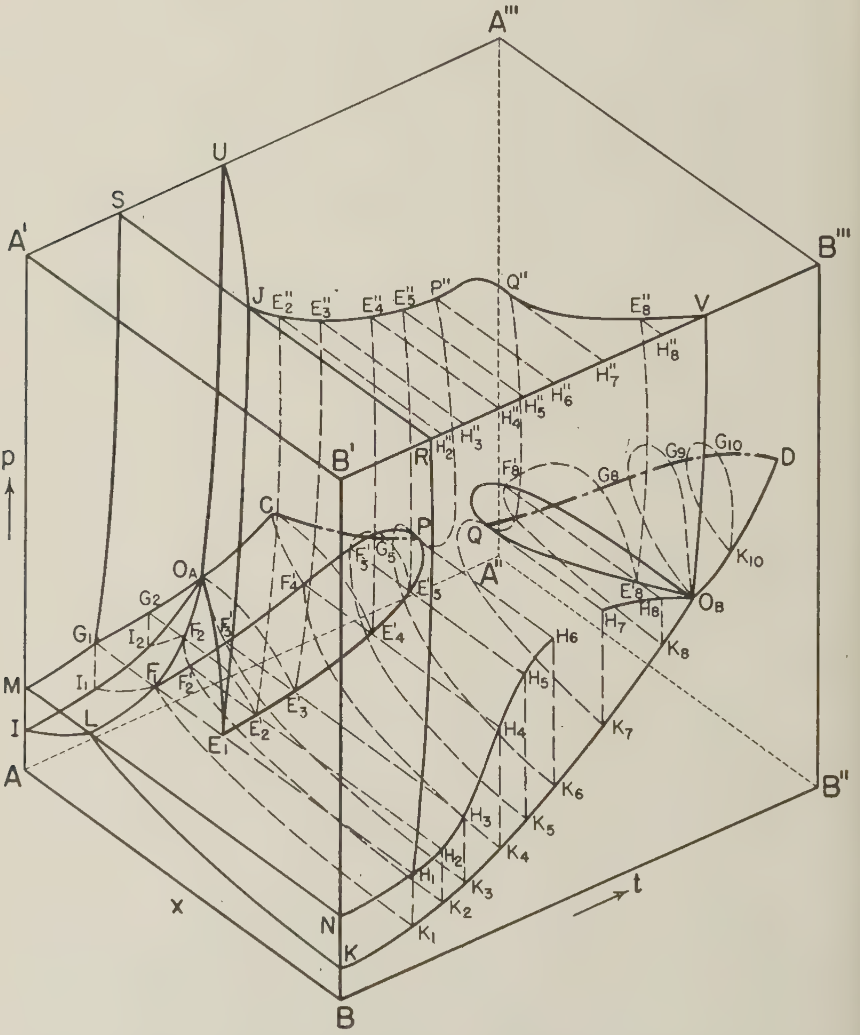

FIG. 1. Isometric drawing of the three dimensional temperature-pressure-concentration diagram of a hypothetical binary system in which there are critical end points and in which the only crystalline phases are the pure components.

Full lines are intersections of the bounding surfaces of fields.

The dash-dot-dash lines are the two segments of the critical curve.

Dashed lines are sections at constant temperature.

In the field $A'B'RSG_1MNH_1$ a random point represents a two phase mixture—crystalline component A and crystalline component B. The proportions of the two phases are given by the relative lengths of the two segments into which a horizontal line parallel to the concentration axis through the random point from the plane $AA'A'''A''$ to the plane $BB'B'''B''$ is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of crystaline A and the segment from the random point to the plane $AA'A'''A''$ represents the mass of crystalline B.

In the field $LIMG_1F_1I_1I_2G_2F_2O_A$ a random point represents a two phase mixture—crystalline component A and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $LIMG_1F_1I_1I_2G_2F_2O_A$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $AA'A'''A''$ represents the mass of the fluid and the segment from the random point to the surface $LII_1F_1F_2I_2O_A$ represents the mass of crystalline component A.

In the field $SJUO_AE_1G_1G_2$ a random point represents a two-phase mixture—crystalline component A and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface JUO_AE_1. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $AA'A'''A''$ represents the mass of the fluid and the segment from the random point to the surface JUO_AE_1 represents the mass of the crystalline component A.

In the field $LKNH_1F_1K_1K_2H_2F_2'F_3'K_3H_3H_4F_4'K_4K_5H_5F_5'PK_6H_6$ a random point represents a two phase mixture—crystalline component B and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $KK_1K_2K_3K_4K_5K_6PF_5'F_4'F_3'F_2'F_1$. The proportions the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of the fluid and the segment from the random point to the surface

$$KK_1K_2K_3K_4K_5K_6PF_5'F_4'F_3'F_2'F_1$$

represents the mass of crystalline component B.

In the field

$$H_1E_1JRH_2''H_2E_2'E_2''E_3''H_3''H_3E_3'E_4'E_4''H_4''H_4H_5E_5'E_5''H_5''H_6''H_6PP''$$

a random point represents a two phase mixture—crystalline component B and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $E_1E_2'E_3'E_4'E_5'PP''E_5''E_4''E_3''E_2''J$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of the fluid and the segment from the random point to the surface $E_1E_2'E_3'E_4'E_5'PP''E_5''E_4''E_3''E_2''J$ represents the mass of crystalline component B.

In the field $K_6H_6''P''PK_7H_7''Q''Q$ a random point represents a two phase mixture—crystalline component B and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $K_6PP''Q''QK_7$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of the fluid and the segment from the random point to the surface $K_6PP''Q''QK_7$ represents the mass of crystalline component B.

In the field $K_7QH_7H_8K_8F_8'O_B$ a random point represents a two phase mixture—crystalline component B and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $K_7QF_8'K_8O_B$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of the fluid and the segment from the random point to the surface $K_7QF_8'K_8O_B$ represents the mass of crystalline component B.

In the field $H_7QQ''H_7''H_8''H_8E_8'E_8''VO_B$ a random point represents a two phase mixture—crystalline component B and fluid. The composition of the fluid is given by the point in which a horizontal line parallel to the concentration axis through the random point intersects the surface $QQ'E_8''E_8'O_BV$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the plane $BB'B'''B''$ represents the mass of the fluid and the segment from the random point to the surface $QQ''E_8''E_8'O_BV$ represents the mass of crystalline component B.

In the field $E_1F_1F_2'F_2E_2'E_3'F_3'O_ACE_4'F_4'F_5'G_5E_5'P$ a random point represents a two phase mixture—fluid I (gas) and fluid II (liquid). The compositions of the two fluids are given by the two points in which a horizontal line parallel to the concentration axis through the random point intersects the bounding surface of the field, the composition of fluid I (gas) being represented by the intersection point nearer the plane $AA'A'''A''$ and the composition of fluid II (liquid) being represented by the intersection point nearer the plane $BB'B'''B''$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line across the field parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the boundary of the field nearer the plane $AA'A'''A''$ represents the mass of fluid II (liquid) and the segment from the random point to the boundary of the field nearer the plane $BB'B'''B''$ represents the mass of fluid I (gas).

In the field $QE_8'F_8'G_8G_9O_BK_{10}G_{10}D$ a random point represents a two phase mixture –fluid I (gas) and fluid II (liquid). The compositions of the two fluids are given by the two points in which a horizontal line parallel to the concentration axis through the random point intersects the bounding surface of the field, the composition of fluid I (gas) being represented by the intersection point nearer the plane $AA'A'''A''$ and the composition of fluid II (liquid) being represented by the intersection point nearer the plane $BB'B'''B''$. The proportions of the two phases are given by the relative lengths of the two segments into which the horizontal line across the field parallel to the concentration axis through the random point is divided by the random point. The segment from the random point to the boundary of the field nearer the plane $AA'A'''A''$ represents the mass of fluid I (gas) and the segment from the random point to the boundary of the field nearer the plane $BB'B'''B''$ represents the mass of fluid II (liquid).

In the remaining field occupying the remainder of the parallelepiped (extending downward from the area $VQ''P''JUA'''B'''$ in the top plane of the parallelepiped and upward from the entire base of the parallelepiped) a random point represents a homogeneous fluid.

In the surface G_1SRH_1 a random point represents a three phase mixture—crystalline component A, crystalline component B and fluid. The composition of the fluid is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve E_1J. The relative masses of the three phases are not determined by the position of the random point (since a particular total composition can be obtained with an infinite number of different mixtures of the three phases).

In the surface MG_1H_1N a random point represents a three phase mixture—crystal line component A, crystalline component B and fluid. The composition of the fluid is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve LF_1. The relative masses of the three phases are not determined by the position of the random point.

In the surface $G_1G_2O_AE_1$ a random point represents a three phase mixture—crystalline component A, fluid I (gas) and fluid II (liquid). The composition of fluid I (gas) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve $F_1F_2O_A$. The composition of fluid II (liquid) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve E_1O_A. The relative masses of the three phases are not determined by the position of the random point.

In the surface $F_1F_2'F_3'F_4'F_5'PH_6H_5H_4H_3H_2H_1$ a random point represents a three phase mixture—crystalline component B, fluid I (gas) and fluid II (liquid). The composition of fluid I (gas) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve $F_1F_2'F_3'F_4'F_5'P$. The composition of fluid II (liquid) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve $E_1E_2'E_3'E_4'E_5'P$. The relative masses of the three phases are not determined by the position of the random point.

In the surface $QF_8'O_BH_8H_7$ a random point represents a three phase mixture—crystalline component B, fluid I (gas) and fluid II (liquid). The composition of fluid I (gas) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve $QF_8'O_B$. The composition of fluid II (liquid) is represented by the point in which a horizontal line parallel to the concentration axis through the random point intersects the curve $QE_8'O_B$. The relative masses of the three phases are not determined by the position of the random point.

A random point on the straight line G_1H_1 represents a four phase mixture—crystalline component A, crystalline component B, fluid I (gas) and fluid II (liquid). The composition of fluid I (gas) is represented by the point F_1. The composition of fluid II (liquid) is represented by the point E_1. The relative masses of the four phases are not determined by the position of the random point on the line G_1H_1.

Diagrams for more complicated types of systems will undoubtedly be needed in the experimental study of systems related to those in which pegmatite dikes and hypogene ore deposits are formed and in the geological application of the experimental results. The two types of binary system in which the only crystalline phases are the pure components serve as stepping stones to the more complicated types.

ACKNOWLEDGMENT

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MELONITE IN BOULDER COUNTY, COLORADO

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Introduction

Melonite, a rare ditelluride of nickel ($NiTe_2$), was first described by Genth (1868), who found minute amounts in a mixture of quartz, hessite, altaite, and, possibly, native tellurium from the Stanislaus mine in the Mother Lode region of California. Later, Hillebrand (1883–1884) identified melonite in telluride ore from the Forlorn Hope mine in Boulder County, Colorado.

Additional occurrences have been noted for the Cripple Creek District of Colorado; Illinawortina, South Australia; and, recently, in Quebec and Ontario, Canada. The Quebec occurrence has been described in detail by Peacock and Thompson (1946), who found that melonite in the Robb-Montbray Mine appears in patches up to 20 mm. wide associated with altaite, tellurbismuth, montbrayite, pyrite, chalcopyrite, and gold, or is disseminated through wall rock. Thompson (1949) reports that small amounts of melonite have been found in the Wright-Hargreaves, Toburn, and Macassa mines at Kirkland Lake in Ontario.

Analyses of concentrates from the telluride ores of certain mines in the Magnolia District in Boulder County consistently yield small amounts of nickel, and, although it has been suspected that the nickel comes from melonite, specimens containing megascopically visible melonite are found only with great difficulty and are virtually nonexistent. In 1946 the writer began a systematic investigation of the telluride ores of the Magnolia District in an effort to isolate the nickel mineral. It soon became apparent that the study should be extended to include all of the telluride ores of Boulder County—a task nearing completion. This paper is intended to describe only one result of this interesting and complex study.

Boulder County Telluride Ores

The Boulder County telluride deposits are at the extreme northeastern end of the mineral belt of Colorado. Much of the ore has been mined from small pockets and lenses in complex, erratically mineralized fissures, many of which trend east or northeast and are related both in space and time to a series of strong northwesterly trending faults, locally called "breccia reefs."

Boulder County has yielded a variety of ores including gold-bearing base metal ores, lead-zinc-silver ores, ferberite (tungsten) ores, and gold-silver telluride ores. The deposits appear to be related genetically to Tertiary igneous activity which began in late Cretaceous or early Tertiary and continued into middle or late Tertiary. In certain mines the telluride ores are associated with either leptothermal lead-zinc-silver or epithermal tungsten mineralization, or both, and field and laboratory studies indicate that all three types of ores have a common origin. In many mines the telluride ores grade downward or laterally into lead-zinc-silver ores and show a crude zonal relationship.

Mineralogically, the telluride ores consist of complex aggregates of chalcedonic quartz, sparse carbonates and sulfates, sulfides, tellurides, and native gold or native tellurium. The mineralogy and structure suggest shallow deposition at a relatively low temperature, and the deposits probably should be classified as epithermal.

Most of the productive veins are found in pre-Cambrian granite, but some ore has been recovered from schist and gneiss, which, locally, has been extensively granitized. The Boulder County telluride ores differ from those in certain other localities, such as Cripple Creek, in that they are not visibly related to extensive extrusive and near surface intrusive igneous activity.

Occurrence of Melonite

Figure 1 shows the salient geologic features of the area of telluride mineralization in Boulder County and the locations of mines yielding ore in which melonite has been identified in the present study. The major late Cretaceous or Tertiary igneous intrusions and the strong northwest trending faults (breccia reefs) are shown in order that the space relationships between these features and the telluride deposits may be grasped. Although specimens from many mines did not yield melonite, it is probable that if a sufficient number and variety of specimens from each mine were available and could be studied, melonite would be found as a constant accompaniment of all or most of the telluride deposits of the area.

Identification of Melonite

A combination of tests was used to establish the presence of melonite in samples of telluride ore. The presence of nickel in gross samples was verified by crushing the ore, dissolving in acid, and applying the dimethylglyoxime test. About a fourth of more than two hundred samples tested in this manner gave a positive test for nickel. However, tests showed that the distribution of melonite in an individual deposit is very erratic. Next, the nickeliferous samples were examined in polished sections, and the melonite was identified by etch reactions and microchemical tests for nickel and tellurium. A final check was made by means of *x*-ray powder techniques.

In polished section the melonite is an early mineral deposited on the walls of small vugs or druses and stands out in relief against gangue minerals and later ore minerals. The mineral generally appears in thin coat-

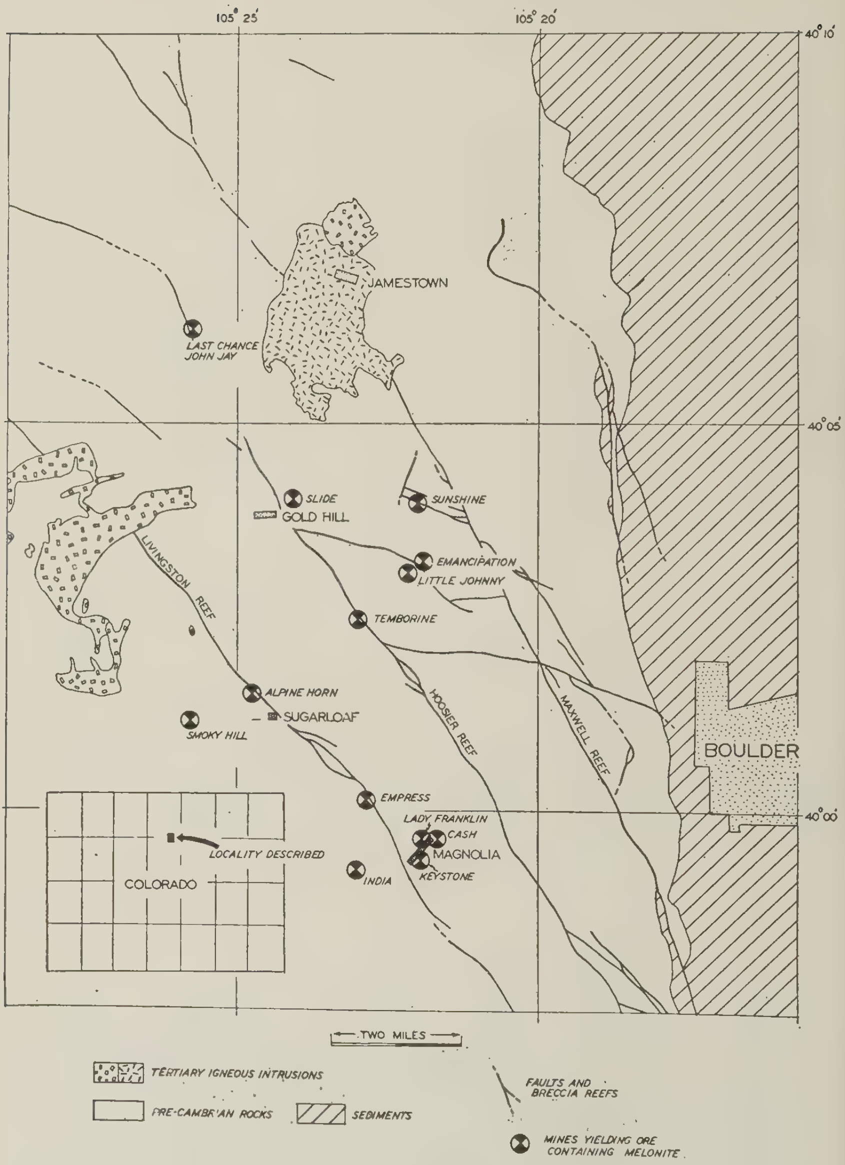

FIG. 1. Outline map of a part of Boulder County, Colorado showing locations of mines yielding melonite.

ings with a colloform structure, the sinuosities of which reflect in detail the irregularities of the walls of the cavities. The general appearance suggests repeated deposition of thin colloidal layers, but under polarized light the melonite is seen to comprise equigranular or radiating aggregates of tiny, moderately to strongly anisotropic grains.

In reflected light the melonite has a pale coppery color, the exact hue of which depends on the color of adjacent, contrasting, gray, white, or yellowish minerals. The etch reactions are consistent with those described by Short (1940) and Thompson (1949). HNO_3 quickly effervesces and stains black; HCl and KCN, negative; $FeCl_3$ slowly stains light to dark brown; KOH and $HgCl_2$, negative.

The *x*-Ray powder patterns gave results that agree in detail with data given by Peacock and Thompson (1946) for melonite from Quebec.

Mineral Association and Paragenesis

Melonite is the earliest telluride mineral in a complex sequence of gangue and ore mineral depositions. It is commonly associated with other later ore minerals and is coated or covered by them. Systematic examination of polished sections of samples from widely scattered deposits in the telluride district reveals a fairly consistent sequence of deposition of gangue and ore minerals. The position of melonite in relation to other minerals in the paragenetic series is shown in Fig. 2. In the

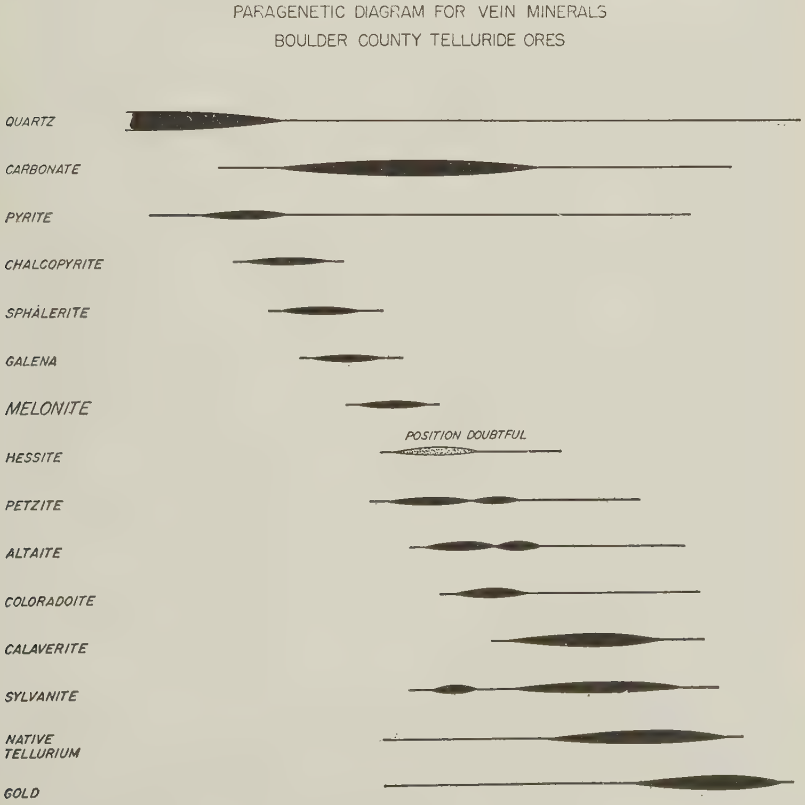

Fig. 2. Paragenetic diagram for Boulder County telluride ores.

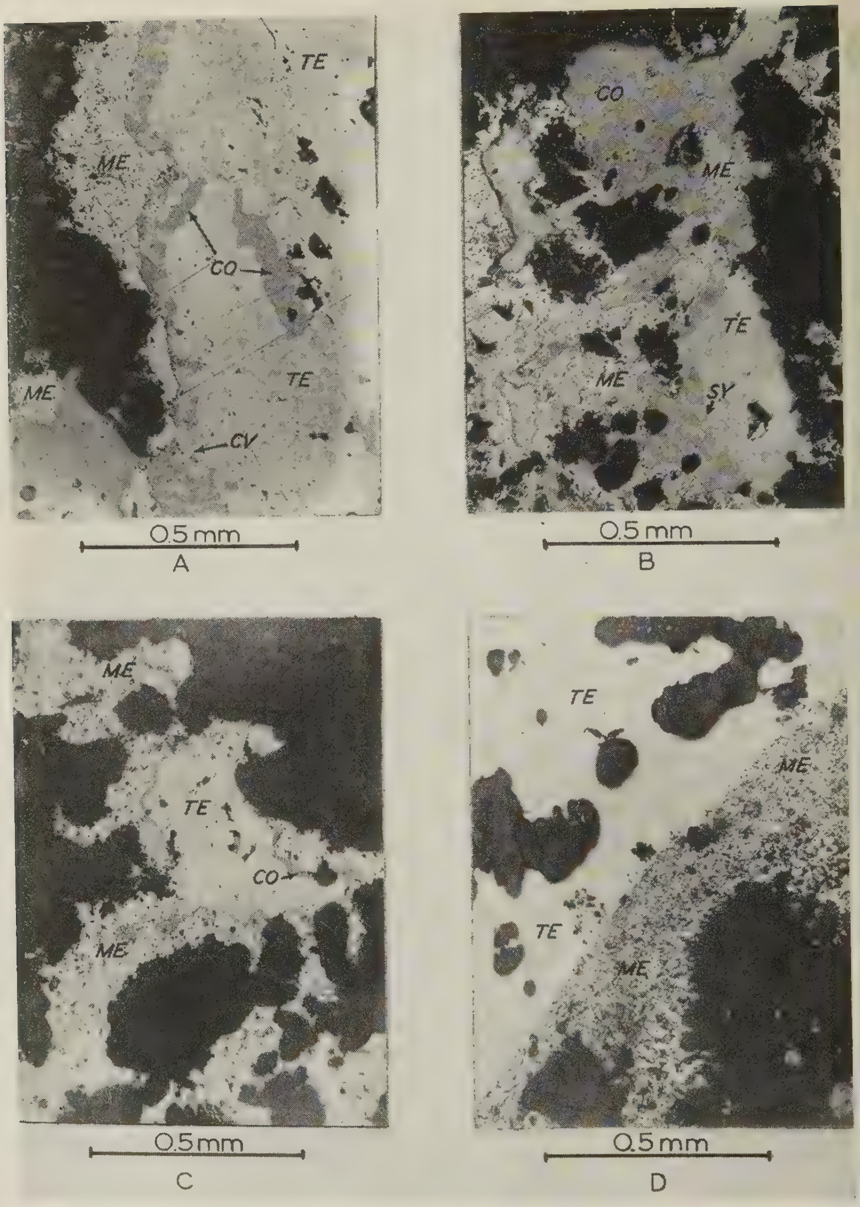

FIG. 3. Photomicrographs of melonite-bearing telluride ores from Boulder County.

A. Melonite (ME), coloradoite (CO), calaverite (CV), and native tellurium (TE). Alpine Horn Mine, Sugarloaf District.

B. Melonite (ME), coloradoite (CO), sylvanite (SY), and native tellurium (TE). Empress Mine, Sugarloaf District.

C. Melonite, coloradoite, and native tellurium. Cash Mine, Magnolia District.

D. Melonite and native tellurium. Sunshine District.

ore mineral sequence melonite follows the sulfides and marks the beginnings of telluride deposition (Fig. 3). Later than the melonite are several tellurides including hessite (Ag_2Te), petzite (Ag_3AuTe_2), altaite (PbTe), coloradoite (HgTe), calaverite ($AuTe_2$), and sylvanite ($AuAgTe_4$). Native tellurium and native gold are the last minerals deposited in the ores, but are not found in proximity to each other in a single ore body.

Although melonite is found with a variety of tellurides, many specimens show a constant association of melonite, coloradoite, calaverite or sylvanite, and native tellurium.

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Much difficulty was experienced in obtaining a representative suite of telluride ores because most of the mines in Boulder County have been inactive for many years. Many of the specimens were obtained from collections at the University of Colorado, but the study did not assume any real significance until owners of private collections became aware of the investigation and generously donated or loaned specimens in their possession. The writer wishes to acknowledge a particular debt of gratitude to Dr. C. N. Eddy, J. E. Byron, and Russell R. McLellan for their contributions of specimens and their continued interest in the progress of the study.

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